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Late Riphean Mineralized Microfossils from the Valyukhta Formation of the Baikal–Patom Highland

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Abstract—New sites of mineralized microfossils, seven in cherts and one in phosphorite of the Upper Riphean Valyukhta Formation of the Baikal–Patom Highland, are discovered at the Alekan River (the Malaya Chuya River basin). The Valyukhta microbiota is specific in taxonomic composition persistent over the area and, thus, significant for biostratigraphy and correlation. It includes four new species of three insertae sedis genera: *Siphonomorpha* Golovenok et Belova, gen. nov.; *Alekania* Golovenok et Belova, gen. nov.; and *Bifaria* Belova, gen. nov., which are described for the first time in this work.

Key words: *Precambrian, Late Riphean, biostratigraphy, correlation, microfossils, microbiota, East Siberia, Patom Highland.*

Single specimens of silicified microfossils from upper cherty beds of the Valyukhta Formation of the Patom Highland were recorded a long time ago. Schopf was the first to distinguish them in a sample of cherty stromatolite from the collection of T.A. Dol'nik (Schopf *et al.*, 1977, 1979). He distinguished only three types of filamentous forms, whose taxonomic position was not considered. Afterwards, we discovered and described the small spiral forms *Obruchevella pussila* Gol. et Bel. from the same site (Golovenok and Belova, 1983).

The systematic investigation of most representative sections of the carbonate Valyukhta Formation and dense sampling of cherts for micropaleontological analysis were performed in 1989–1991 under the guidance of V.K. Golovenok. The studied sections are located along the Bol'shoi Patom River upstream off its junction with the Lena River, along the Vitim River and its right-hand tributaries (Pudrikha and Bystraya rivers, and along the Malaya Kalaika and Alekan rivers in the Malaya Chuya River basin. In addition, we studied the chert samples from the Sen' Formation collected by N.K. Korobeinikov in the Zhuya River basin (Fig. 1) and the fine-grained phosphorite samples collected by V.K. Golovenok in upper courses of the Alekan River.

The sections are studied and sampled for cherts to a variable extent. The micropaleontological analysis of thin sections revealed, however, that microfossils occur in all of them (Fig. 2).

The distinguished Valyukhta microbiota is not very diverse but very specific in the taxonomic composition. The latter is persistent in all sites, and the identified microfossil taxa are of a high significance for biostratigraphic correlation.

STRATIGRAPHIC POSITION, DISTRIBUTION AREA, AND LITHOSTRATIGRAPHY OF THE VALYUKHTA FORMATION

The Dzhemukan, Barakun, and Valyukhta formations are constituents of the Dal'nyaya Taiga Horizon (Group) of the Patom Highland, where it remains disputable with respect to the stratigraphic age. According to resolutions of the Stratigraphic Conference on Precambrian, Paleozoic, and Quaternary Deposits of Central Siberia (Novosibirsk, 1979), the Middle–Upper Riphean boundary is "conventionally placed inside the horizon, and principles to recognize it in the region are equivocal" (*Resheniya ...*, 1983, p. 30).

Despite this and the lack of radiometric dates and paleontological constraints, many researchers are quite sure that the Valyukhta Formation proper is Late Riphean in age. Carbonate rocks of the formation include stromatolites *Baicalia valuchtenia* Dol. and *B. hirta* Dol., and the Sen' Formation of the southeastern area of the highland, which is presumably correlative to these rocks, yielded *Baicalia nitchatica* Dol., *B. tcharica* Dol., *Conophyton garganicus* Korol., and *C. metula* Kir. (*Resheniya ...*, 1983). Both the *Baicalia* and *Conophyton* forms are known to occur not only in the Middle, but also in the Upper Riphean.

Stanevich and Faizulina (1992) recently described the organic-walled microfossils from clay deposits of the Valyukhta Formation. The assemblage of these remains includes taxa of wide stratigraphic ranges: *Protosphaeridium* (Tim.) Stan., *Leiosphaeridia* (Eisenack) Downie et Sarjeant, *Synsphaeridium* Eisenack,

† Deceased.

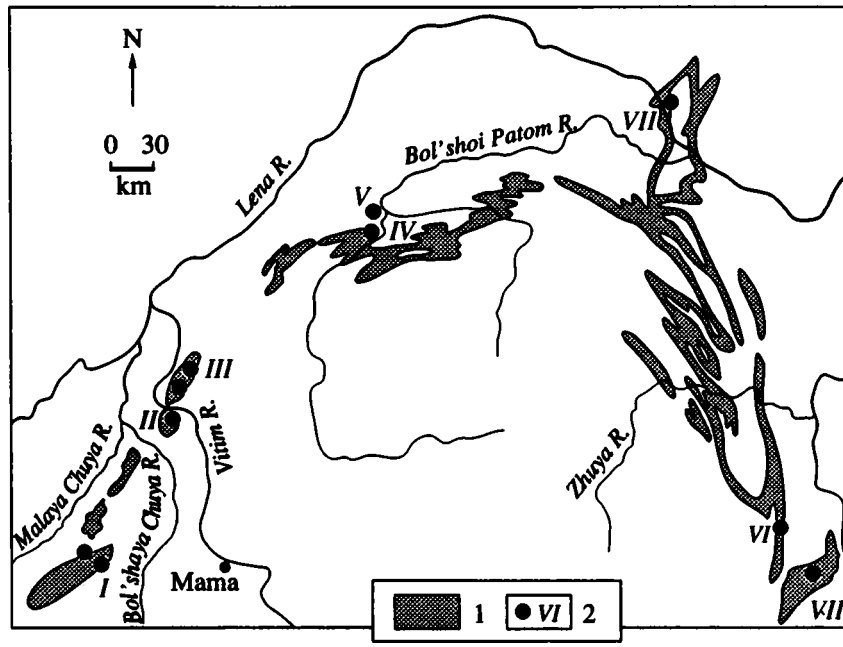


Fig. 1. Localities of microfossils recorded in the Valyukhta (*I–V, VII*) and Sen' (*VI, VIII*) formations of the Patom Highland. Distribution areas of the Valyukhta and Sen' formations (1) and studied localities (2) of microfossils: *I*, Malaya Chuya River basin (Malaya Kalaika and Alekan rivers); *II*, Vitim River; *III*, upper courses of the Pudrikha and Bystraya rivers; *IV*, Bol'shoi Patom River upstream from the Chelonchen River mouth; *V*, ditto, downstream from the Chelonchen River mouth; VI, Bestyakh River; *VII*, left-hand bank of the Lena River counter to Tinninskii Is.; *VIII*, Kumakh-Ulakh Creek, Zhuya River basin.

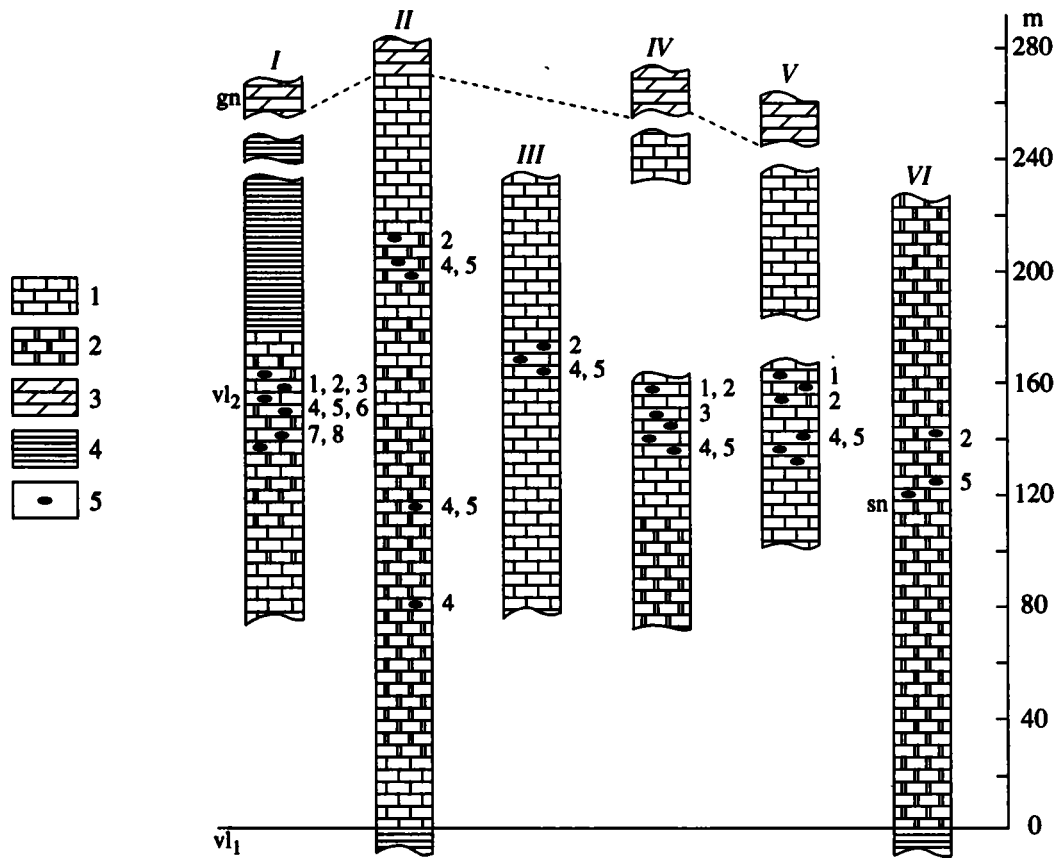


Fig. 2. Schematic lithologic columns of the upper Valyukhta Subformation (*I–V*) and Sen' Formation (*VI*): (1) limestone; (2) dolomite; (3) marl; (4) terrigenous rocks; (5) cherty lentils; (vl_1, vl_2) lower and upper Valyukhta subformations; (gn) Zherba Formation; (sn) Sen' Formation; Roman numbers correspond to localities indicated in Fig. 1 and Arabic numerals denote positions of microfossil taxa listed in Fig. 3.

Symplassosphaeridium Tim., and others associated with endemic forms of genera *Rosella* Stan., *Radiata* Stan., *Proteus* Stan., and *Nucellosphaera* Stan., amid which there could be present some artifacts (Golovenok and Belova, 1995).

Thus, an important objective of regional stratigraphy is to obtain convincing paleontological data substantiating the Late Riphean age of the Valyukhta Formation.

Deposits of the Valyukhta Formation are widespread in the Lena River fold zone, which is an extent band of outcrops of Upper Proterozoic to Lower Paleozoic rocks occurring between the Siberian platform and an arcuate chain of the Chuya, Tonod, and Longdor anticlinoria representing inner parts of the Patom and North Baikal highlands (Golovenok, 1976). This huge (over 700 km long) zone extends through the middle–lower courses of the Bol'shoi Patom River from the Chuya River in the southwest to the Nichatka Lake in the southeast. Salop (1964) attributed this territory to the Lena River zone of the Late Proterozoic Baikal–Patom depression ending in the northeast with the Ur lateral branch (Fig. 1).

Within the Lena River fold zone, Riphean and Vendian formations are highly variable with respect to lithofacies characteristics and thickness, especially in its southwestern and southeastern limbs, where the total thickness of deposits is extremely decreased (Golovenok, 1976). This is reason why our colleagues from Irkutsk divide the Lena River zone into the individual Chuya, Chelonchen–Zhuya, Ur, and Nichatka subzones, as it is registered in resolutions of the aforementioned stratigraphic conference (*Resheniya* ..., 1983).

The Valyukhta Formation conformably rests on carbonates of the Barakun Formation and consists of gray to dark-gray and black carbonaceous shales, siltstones, sandstones, limestones and dolomites. Proportions between these rocks considerably vary from one section to another. It is a rather common practice to divide the Valyukhta Formation into two–three subformations, although in a conventional manner. In the Ur subzone, the respective subdivisions were even authorized as a succession of individual formations (*Resheniya* ..., 1983). The total thickness of the Valyukhta Formation, most considerable in the north inside the Chelonchen–Zhuya subzone, is within the range of 1000 to 1600–1800 m.

There is a general regular trend in lithologic and structural changes of the Valyukhta Formation; these changes are most distinct in the Chelonchen–Zhuya and Ur subzones, where its rocks are exposed wider than elsewhere. In inner sections of the Patom Highland, this formation predominantly consists of terrigenous rocks. Its sections located to the north and northeast, closer to Siberian platform, demonstrate sequentially growing proportions of carbonate deposits represented by gray to dark-gray limestones and dolomites. Near the platform, these rocks compose a separate, rather thick (up to 300 m) member at the top of the formation. This trend also seems to be characteristic of the southwestern Chuya subzone of the Lena River fold arc.

The situation is more complex in the southeastern Nichatka structural subzone, where the uppermost carbonate member is missing in the northwestern sections of the so-called Sharapovo syncline. To the south and southeast from here, however, this interval of the Riphean sequence is occupied by the Sen' Formation of dominant dolomite composition resting on a subordinate basal sandstone unit. The last formation is usually thought to be correlative to the Valyukhta Formation, although direct relations between both have never been observed.

In the eastern submeridional branch of the Chelonchen–Zhuya structural subzone, the lithology of the Valyukhta Formation is different. Here, beginning from the lower courses of the Bol'shoi Patom River in the north to the Zhuya River in the south, its topmost thick unit (up to 300–500 m) consists of dark-gray sandstone and siltstone beds, whose matrix, however, is calcareous as a rule.

The upper limestone and dolomite beds of the Valyukhta Formation are gray to dark-gray and black. Lighter colors are more characteristic of the second rock type. Many of these carbonate beds display the oncolite-detrital texture and massive or locally bedded structure. In addition to small oncolites (few fractions of millimeter in size), some beds include large pisolites exceeding 2 cm and more in diameter. Carbonate clastics also vary in size from sandy grains to flattened pebbles up to few centimeters long.

In all sections located near the platform, upper carbonate beds of the Valyukhta Formation enclose lenses and lenticular interbeds of cherts. They are up to 5–7 cm in thickness and 0.2–1 m in length. Cherts are brown- to dark-gray or black and display the oolite-clastic texture under the microscope. Microfossils frequently occur in the clastic fraction, occasionally in matrix, and never in layered oncolite envelopes.

TAXONOMY, DISTRIBUTION, AND BIOSTRATIGRAPHIC SIGNIFICANCE OF VALYUKHTA MICROBIOTA

Taxonomic composition of the studied microbiota and distribution patterns of microfossils in sections are illustrated in Figs. 2 and 3. In our synopsis, we do not consider data by Stanevich and Faizulina (1992) on organic-walled microfossils and our own data on sites of the Lena River and Kumakh-Ulakh Creek (Sen' Formation of the Zhuya River basin), where we identified only long-ranged forms of *Siphonophycus* (Schopf) Knoll *et al.*

Most sections of the Valyukhta Formation yielded filamentous microfossils representing a characteristic assemblage of *Siphonophycus robustus–Siphonophy-*

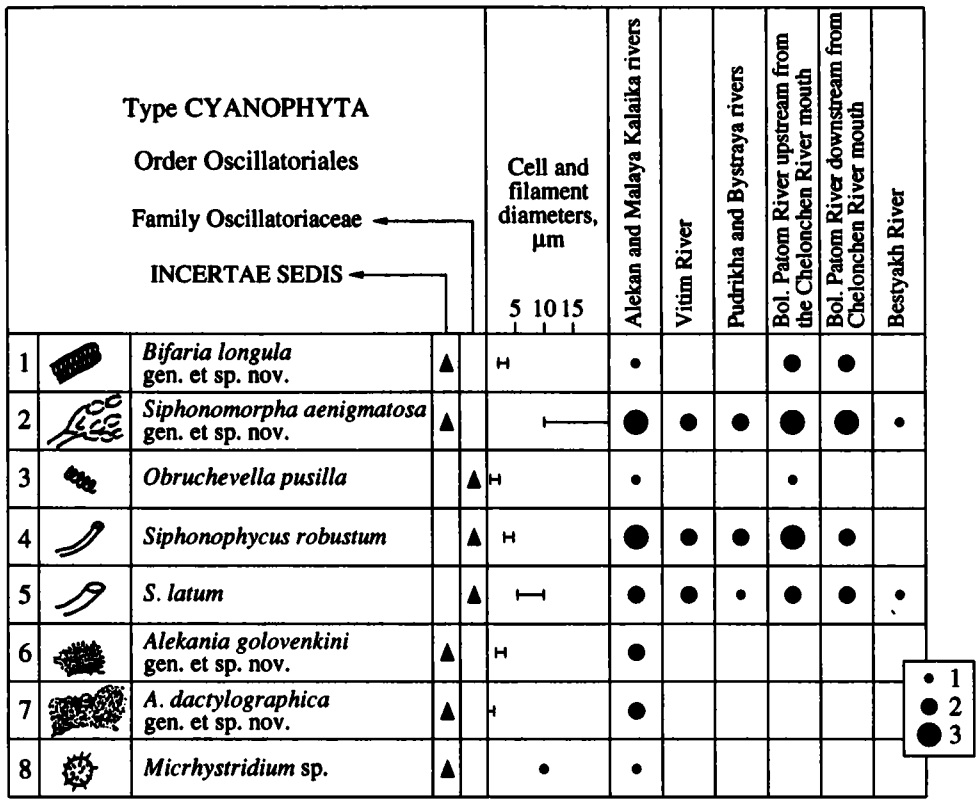


Fig. 3. Synopsis of the Valyukhta microbiota: (1) single specimens; (2) more than 10 specimens; (3) dozens and hundreds of specimens.

cus latum–Siphonomorpha aenigmatosa gen. et sp. nov. The first two species of this assemblage are of a wide geographic and stratigraphic range, whereas *S. aenigmatosa* is detected so far only in the Valyukhta and Sen' formations. Therefore, the assemblage is not suitable for interregnal correlation. At the same time, the peculiar morphology and abundance of relatively large *S. aenigmatosa* forms represent quite distinctive features of this microbiota traceable over a wide area of the Patom Highland, and the assemblage can be successfully used for correlation of the Valyukhta and Sen' formations in the region.

The same, though to a lesser extent, is characteristic of specific taxon *Bifaria longula* gen. et sp. nov., unfortunately missing in some sections (Fig. 3). The last fact may be a consequence of two factors: first, these microorganisms might be very scarce and irregularly dispersed; second, one should bear in mind that some sections at the Pudrikha, Bystraya, and Bestyakh rivers are inadequately studied.

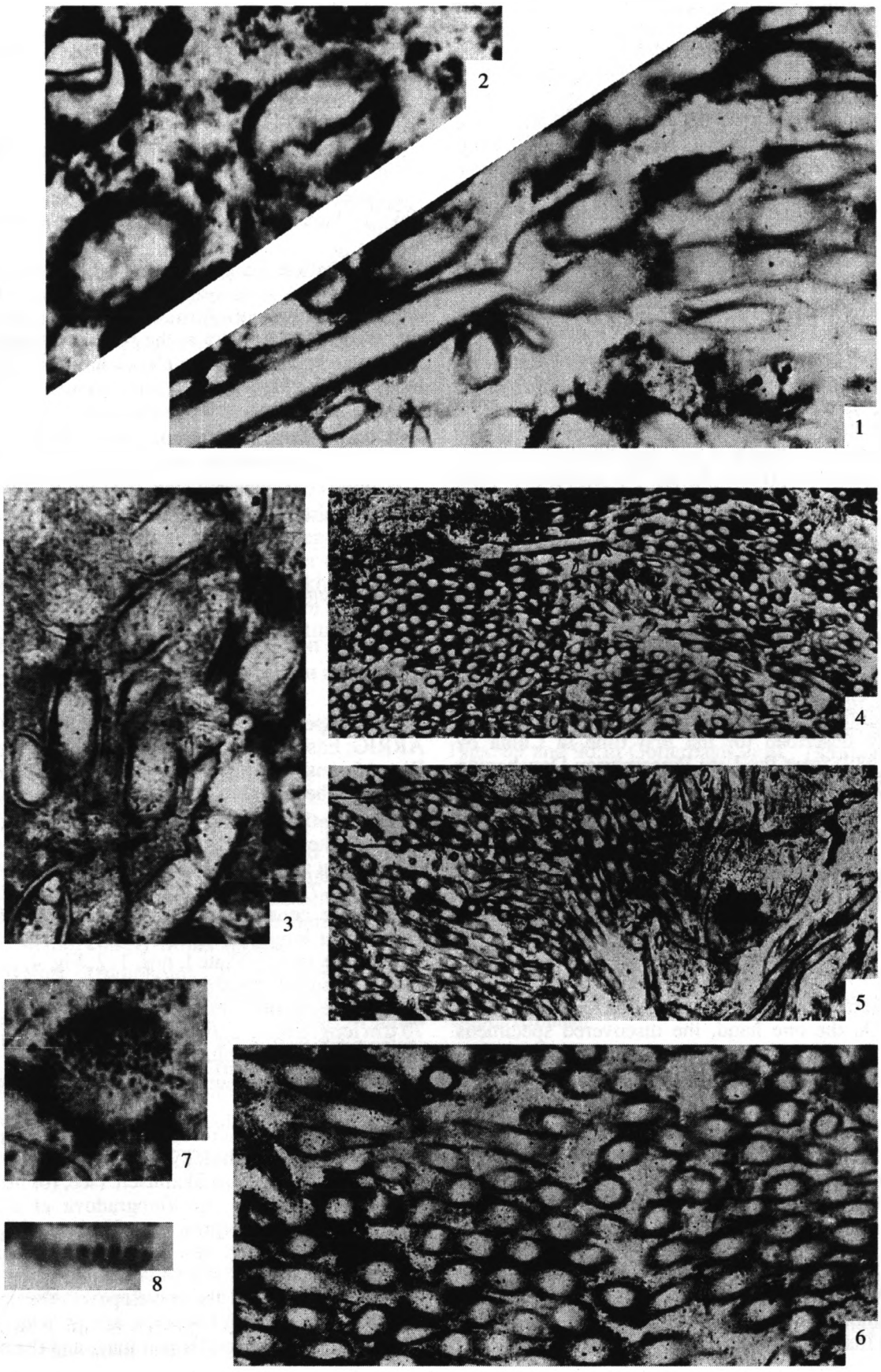
Small spiral species *Obruchevella pussila* Gol. et Bel. are occasional components of the Valyukhta microbiota (Golovenok and Belova, 1983). In addition to their type locality at the Bol'shoi Patom River, we discovered them also at the Alekan River site (Malaya Chuya River basin) remote from the former for more than 300 km (Plate I, no. 8). Spiral forms of this type occur in Precambrian deposits of Eurasia, Greenland, and Australia beginning from the Late Riphean level, where they are relatively rare. The Vendian interval corresponds to the thriving period of *Obruchevella* forms: their occurrence sites are much more frequent at this level, the species become larger in size, and their "giant" forms missing in younger deposits finally appear at the Vendian–Cambrian boundary interval (Golovenok and Belova, 1995). All this suggests a high biostratigraphic potential of the genus *Obruchevella* Reitl. applicable for interregional correlation and age determinations. Very small and scarce *O. pussila* forms reported from two widely spaced outcrops of the Valyukhta deposits suggest that this formation is the Late Riphean in age.

Thus, all taxa characteristic of the Valyukhta biota and occurring at least in two sites can be tentatively

Plate I. Mineralized microfossils from cherts of the Valyukhta Formation

(1–6) *Siphonomorpha aenigmatosa* gen. et sp. nov.: (1) holotype, thin section no. 418-b-10-1, ×250; (2) transverse sections of filaments, the same thin section, ×600; (3) thin section no. A-13, ×300; (4, 5) thin section no. 418-b-1, ×50; (6) the last thin section, ×100; all specimens are from the type locality at the Bol'shoi Patom River downstream from the Chelonchen River mouth; (7) *Micrhystridium* sp.; (8) *Obruchevella pusilla* Gol. et Bel.; thin section 487-2, magn. 1000, the Alekan River.

Plate I



divided into three categories of different biostratigraphic significance: (1) species *Obruchevella pusilla* Gol. et Bel. displaying the interregional correlation potential and useful for age determination of host deposits; (2) species *Siphonomorpha aenigmatosa* gen. et sp. nov. and *Bifaria longula* gen. et sp. nov., both characteristic of the Valyukhta and Sen' formations and applicable for the regional correlation; and (3) long-ranged taxa *Siphonophycus* div. sp. useless in the biostratigraphic aspect. The assemblage of organic-walled microfossils that was reported by Stanevich and Faizulina (1992) from terrigenous deposits of the Valyukhta Formation at sites of the Zhuya, Bol'shoi Patom, and Bol'shaya Chuya rivers consists predominantly of sphaeromorphic acritarchs demonstrating wide stratigraphic ranges.

In our synopsis (Fig. 3), there is one more form detected at the Alekan River site only. Regretfully, this site is poorly studied and represented in our collection by several chert and black phosphorite samples only. In these cherts, we identified one specimen of acanthomorphic acritarch (Plate I, no. 7) associated with *Obruchevella* forms and other characteristic taxa of the Valyukhta microbiota. This specimen of spheroid envelope 14 μm in diameter and lacking the external membrane is ornamented with short tiny (1–2 μm) processes and can be attributed quite positively to the genus *Micrhystridium* (Deflandre) Downie et Sarjeant. Apparently, this is one of the most ancient forms of the genus.

Microfossils that have been detected in thin sections of calcareous-sandy phosphorite are especially interesting. Mineralized microfossiles from the phosphorite beds were described for the first time in China by Zhang (1989) from the Late Precambrian Doushantuo Formation. He classed them with red algae taking into account their large size and differentiated type of thalluses supplemented with specialized reproductive and vegetative structures. Thalloid forms occur as well in our phosphorite samples, but they are lesser in size and lacking differentiated cells and specialized structures. We classified them as microfossils representing new genus *Alekania* Golovenok et Belova, whose taxonomic position is uncertain.

Occurrence of this genus is interesting in two aspects. On the one hand, the discovered specimens represent an unusual type of microfossils, and, on the other, it confirmes that, besides the cherty concretions, the Precambrian phosphorites are promising objects for further micropaleontological studies.

New taxa of microfossils from the Valyukhta Formation are described below. Preliminary microscopic investigation of thin sections and analysis of taxonomic affinity of microfossils were performed in cooperation with the late V.K. Golovenok, who suggested new generic terms *Siphonomorpha* and *Alekania*. All thin sections with microfossils are stored at the Department of Precambrian Geology (DPG) of the All-Russia Research Institute of Geology (ARRIG).

SYSTEMATIC DESCRIPTION

Incertae sedis

Genus *Siphonomorpha* Golovenoc et Belova, gen. et sp. nov.

Type species: *Siphonomorpha aenigmatosa* Belova, gen. et sp. nov.

Diagnosis. Filaments are rather thick, aseptate and cylindrical in their lower part, but amply branching in the upper. Branches consist of short segments. Envelope is lamellose. Interlacing filaments form dense mats.

Description. Large cylindrical filaments are amply branching and racemose in the upper part. Branches consist of short dolioform to oval and rounded segments (cells ?). Locally at the point of branching, filaments are constricted. Envelopes are thick and lamellose. Usually, filaments are constituents of mats; in the isolated state, they occur sporadically.

Comparison. New genus has nothing in common with other genera of Precambrian microfossils.

Genus composition. Type species.

Distribution. The Upper Riphean of the Patom Highland.

Siphonomorpha aenigmatosa Belova, gen. et sp. nov.

Plate 1, nos. 1–6; Fig. 4, drawings *a–d* and *h*

Species name: from *aenigmatosus* (Latin)—mysterious.

Holotype. Thin section no. 417-b-10-1, DPG ARRIG; East Siberia, Patom Highland, Bol'shoi Patom River downstream from the Chelonchen River mouth; Upper Riphean, Valyukhta Formation.

Description. Empty cylindrical filaments dichotomous in their upper part are 20–55 μm in diameter (32.3 μm in average, n =359) and have bi- to triploblastic walls 2–7 μm thick (3.9 μm in average, n = 359). Filaments are constricted at some points of branching, where their longitudinal and transverse sections show the oblique septae (Plate I, nos. 1, 2; Fig. 4, drawings *c* and *d*). Branches consist of short dolioform to oval and rounded segments (cells ?) 20–55 μm wide and 20–70 μm long. Septae between segments are missing, and their boundaries are blurred.

Comparison. New species alone represents the genus.

Remarks. The microfossils under consideration are most similar in morphology to the recent green algae of the genus *Gemmiphora* Skabitsch. (see, for instance, *G. compacta* Skabitsch. in Vinogradova *et al.*, 1980). Thallus of the latter represents a filament with the racemose upper end (Fig. 4e). Branching of *Gemmiphora* species is peculiar: it begins at the site where a cell of filament grows into the envelope of the lower cell (Figs. 4f, 4g). Their filaments are as wide as 11 to 40 μm, cells are up to 207 μm long, and the envelopes

are thick and lamellose. Each mat consists of a dense fascicle of filaments. All these features of the recent green alga, the size parameters inclusive, are more or less recognizable in the species of *Siphonomorpha aenigmatosa*. At the same time, fossil specimens do not display the distinct cellular structure and reproductive organs (spores, akinetes) or cell nuclei typical of eucaryotes. It also unclear whether the ancient alga was attached like its modern analogue.

Material. We managed to detect only one complete thallus of *Siphonomorpha aenigmatosa* representing the cylindrical filament clearly displaying branching and racemose upper termination (Plate I, nos. 1, 4). In thin sections, there are usually seen aggregates of segments suggesting little about the thallus morphology (Plate I, nos. 3–6), fragments of thick-walled cylindrical filaments up to 300 μm long (Plate I, no. 5), and their transverse sections (Plate I, no. 2). The studied 53 thin sections of 46 chert samples yield more than 1000 fragments of thalluses. Amid them, more than 300 specimens are well preserved.

Distribution in the Patom Highland of East Siberia: Bol'shoi Patom River near the Chelonchen River mouth, Malaya Kalaika and Alekan rivers (Malaya Chuya River basin), Vitim River, and upper courses of the Pudrikha and Bystraya rivers— Valyukhta Formation of the Upper Riphean; middle courses of the Bestyakh River—Sen' Formation of the Upper Riphean.

Genus *Bifaria* Belova, gen. nov.

Type species: *Bifaria longula* Belova, gen. et sp. nov.

Diagnosis. Filamentous forms lacking branching and composed of two rows of closely-spaced oval and rounded cells enclosed into a sheath.

Description. Oval to locally rounded cells are arranged into a couple of parallel closely spaced rows. Cells are equal in size all along the filament. Longer axes of oval cells are oriented across the filament. Each filament has a thin closely-fitting sheath.

Comparison. In morphology, the new genus is most close to the genus *Polysphaeroides* Hermann. This Late Precambrian filamentous microfossils consists of multiserial clusters of spheroid cells. In contrast, specimens of *Bifaria* gen. nov. are of the distinct biserial structure and have thin closely-fitting sheaths.

Composition. Type species only.

Distribution. Upper Riphean of the Patom Highland.

Bifaria longula Belova, gen. et sp. nov.

Plate II, nos. 5 and 6; Figs. 4h and 4i

Species name: from *longulus* (Latin)—elongated.

Holotype. Thin section no. A-13, DPG ARRIG; East Siberia, Patom Highland, Bol'shoi Patom River downstream from the Chelonchen River mouth; Valyukhta Formation of the Upper Riphean.

Description. Oval to locally rounded cells are from 2×3 to 3×3.5 μm in size (2.7×3.4 μm in average, $n = 272$); they are densely arranged into two parallel closely-spaced series forming filaments 20–30 to 170 μm long. Longer axes of cells are oriented across the filament. Filaments are rectilinear or arcuate, enclosed into thin (about 1 μm) well-fitting sheaths, some of which are poorly visible.

Comparison. At present, we know only one species of the genus.

Remarks. Microfossils under consideration are similar in morphology to some cyanophyte species from the genus *Chamaesiphon* Br. et Grun. Exospores of these species occasionally are represented by filamentous biserial aggregates (see, for instance, *Ch. curvatus* f. *polysporinus* Schirsch. in Gollerbakh *et al.*, 1953). In the type site, we identified the sole specimen of *Bifaria longula* almost identical to exospores of *Chamaesiphon* cyanophytes. In the thin section, this specimen is unfortunately located against the background of brown organic matter disguising it in photographs. At the same time, the specimen is well seen under the microscope when enlarged to 200 times or more. One can see in the drawing (Fig. 4j) that the hollow ringed filament is a little wider in the upper part and consists of two series of small oval cells. One end of the filament is concealed, and another seems to be detached. The ringed shape was hardly characteristic of the fossil during the life time, and its similarity with exospores of *Chamaesiphon* forms is quite evident (see Figs. 89 and 90 in Gollerbakh *et al.*, 1953). However, we did not observe other stages of development of this microorganism, and the degree of identity between *Bifaria longula* and *Chamaesiphon* forms remains unknown.

Material. A few dozen filament fragments detected in three thin sections of three rock samples; twenty specimens are well-preserved.

Localities. The Patom Highland of East Siberia: Bol'shoi Patom River near the Chelonchen River mouth and Alekan River; Valyukhta Formation of the Upper Riphean.

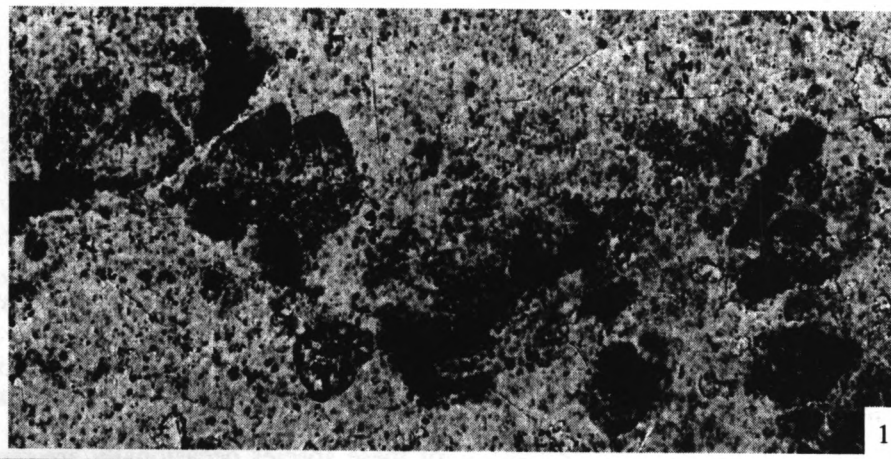
Genus *Alekania* Golovenoc et Belova, gen. et sp. nov.

Type species: *Alekania golovenkini* Belova, gen. et sp. nov.

Diagnosis. Thalluses diverse in morphology consist of plentiful small cells arranged into blurry series.

Description. Tubercular thalluses of irregular flabelliform to hemispheric shape consist of populations of elongated (oval to bacillary) and round cells arranged into blurry, seemingly branching series. The inner cell differentiation has not been observed and reproductive organs are unknown.

Plate II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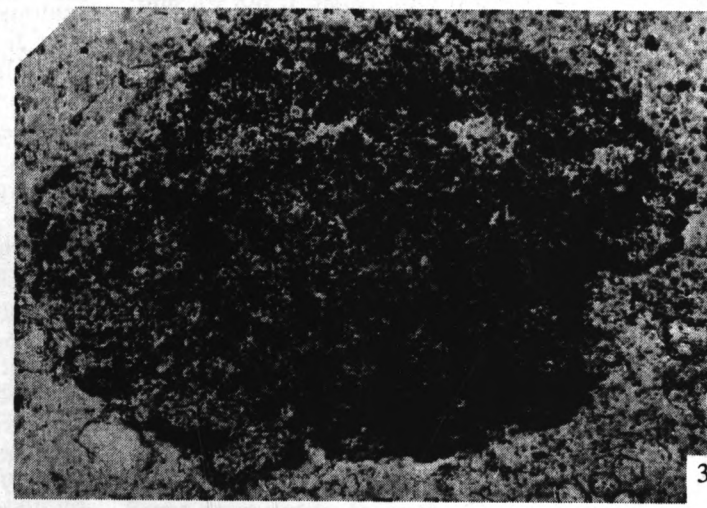
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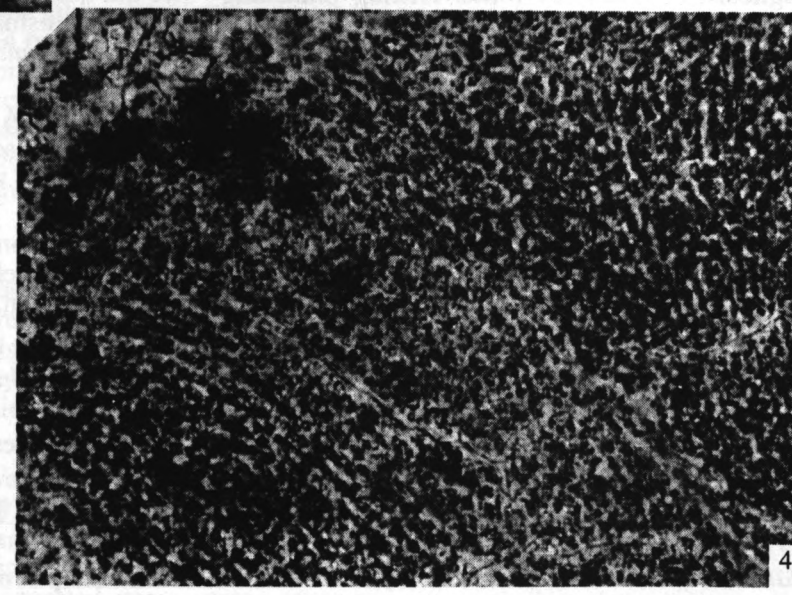
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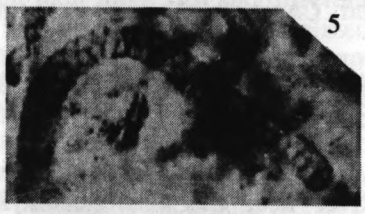
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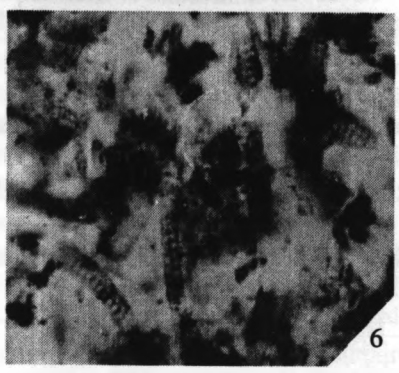
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Comparison. In general morphology, the *Alekania* forms are similar to thalluses belonging to the genus *Thallophyca* Zhang discovered, like the Valyukhta specimens, in phosphorite concretions from the Doushantuo Formation of China (Zhang, 1989). However, the thalluses in question and their cells are lesser in size. In addition, the latter do not display the cell differentiation characteristic of the *Thallophyca* thalluses, and this discredits the affinity of microfossils with the last genus.

Remarks. The structure of the *Alekania* thalluses resembles that of recent entophysaledalean cyanophytes from the genus *Chlorogloea* Wille, whose thalluses display the blurry cell series.

Distribution. Upper Riphean of the Patom Highland.

Composition. Species *Alekania golovenkini* and *A. dactylographica.*

Alekania golovenkini Belova, gen. et sp. nov.

Plate II, nos. 1–4

Species name is given in honor of V.K. Golovenok.

Holotype. Thin section no. 487-5, DPG ARRIG; East Siberia, Patom Highland, Alekan River; Valyukhta Formation of the Upper Riphean.

Description. Thalluses are irregular in shape, tubercular, flabelliform to hemispheric. It is likely that the diverse morphology partially depends on the orientation of thalluses relative to the thin-section plane. Size parameters (width and height) are in the range from a few dozen to 300 μm. Thalluses consist of elongated to round cells usually arranged into blurry radial, presumably, branching series (Plate II, nos. 2 and 4). Oval cells vary in size from 1.2 × 1.5 to 3.0 × 3.5 μm, bacillary from 1.2 × 2.5 to 3.0 × 6.0 μm, and round cells are 3.0–3.5 μm in diameter. Thalluses are composed of one-type cells, whose distribution density inside the thallus is variable.

Comparison. The species differs from *A. dactylographica* sp. nov. in the shape of thalluses, which are lesser in size and show the radial arrangement of cell series.

Material. More than twenty thalluses detected in two thin sections of the calcareous-sandy phosphorite.

Locality. Identical to that of holotype.

Alekania dactylographica Belova, gen. et sp. nov.

Plate III, nos. 1–3

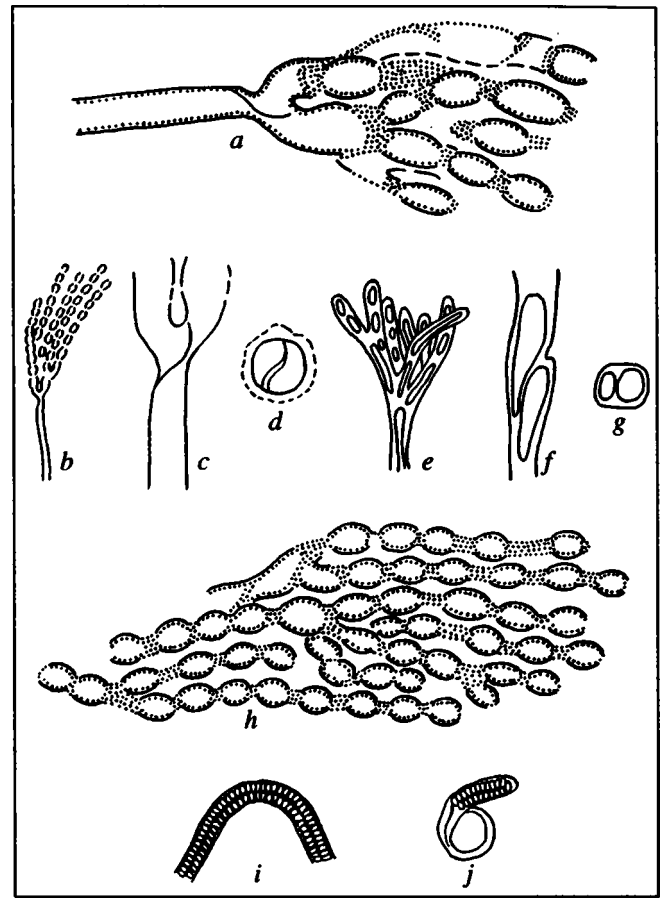


Fig. 4. Sketches illustrating characteristic morphological features of *Siphonomorpha aenigmatosa* gen. et sp. nov. (*a–d, h*) and *Bifaria longula* gen. et sp. nov. (*i, j*) as compared to the thallus of recent green alga *Gemmiphora compacta* Skabitsch (*e*) and its longitudinal (*f*) and transverse (*g*) sections redrawn from Vinogradova *et al.* (1980, p. 25, Fig. 1).

Species name: from *dactylos* (Greek)—finger—and *graphica* (Latin)—image.

Holotype. Thin section no. 487-5-1, DPG ARRIG; East Siberia, Patom Highland, Alekan River; Valyukhta Formation of the Upper Riphean.

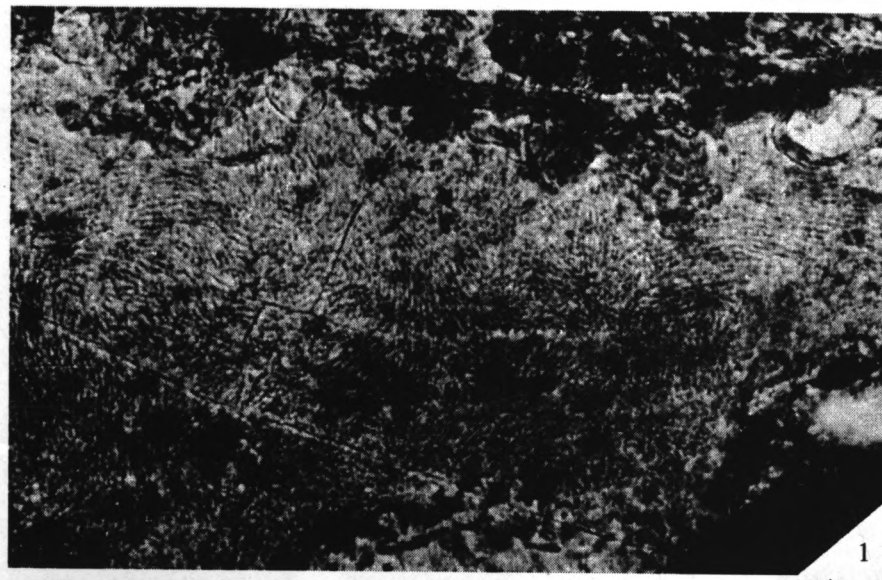
Description. Wide (up to 500 μm across) thalluses of irregular shape consist of elongated bacillary cells arranged in series; the series are winding to capriciously curved, so that the general patterns resemble fingerprints. Cells are 0.8–1.2 μm wide and 2.0–6.2 μm long.

Plate II. Mineralized microfossils from phosphorite (nos. 1–4) and cherty (nos. 5, 6) concretions of the Valyukhta Formation

(1–4) *Alekania golovenkini* gen. et sp. nov.: (1) thin section no. 487-5-II, ×100; (2) holotype, thin section no. 487-5-IV, ×600; (3) thin section no. 487-5-II, ×200; (4) thin section no. 487-5-IV, ×600; all specimens from the phosphorite sample collected in upper courses of the Alekan River.

(5,6) *Bifaria longula* gen. et sp. nov.: (5) holotype, thin section no. A-13, ×600; (6) aggregate of fragmented filaments, thin section no. 418-b-10-1, ×300; all specimens from the type locality at the left-hand bank of the Bol'shoi Patom River downstream from the Chelonchen River mouth.

Plate III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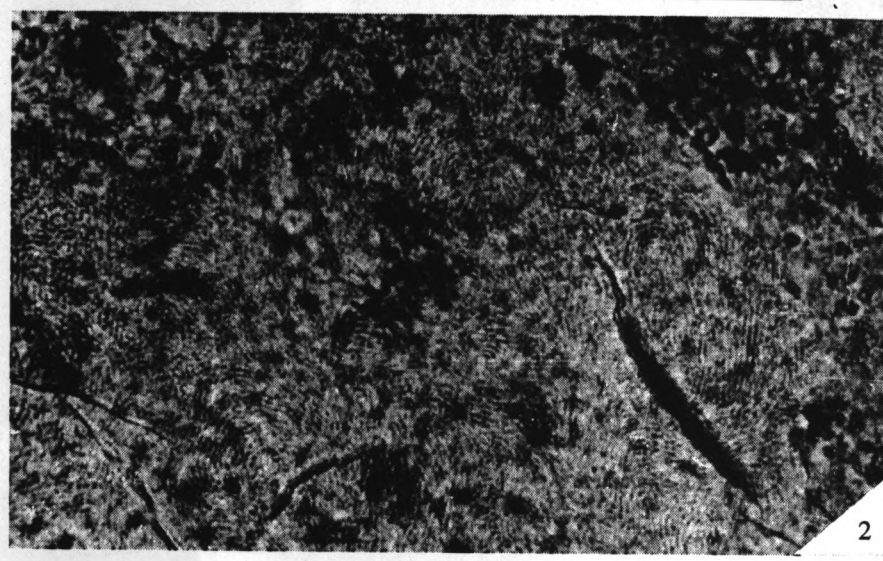
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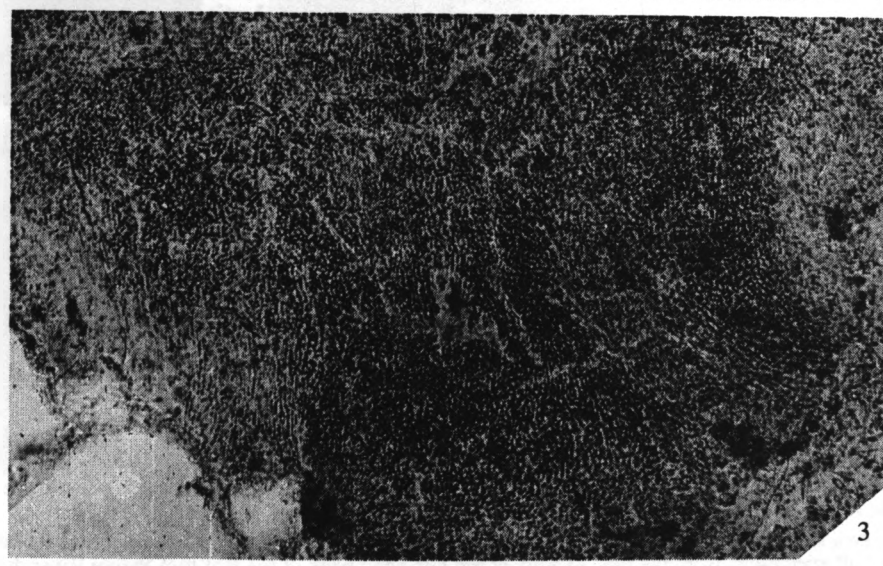
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Plate III. Mineralized microfossils from phosphorite concretion of the Valyukhta Formation.
(1–3) *Alekania dactylographica* gen. et sp. nov.; holotype (1) and other specimens, thin section no. 487-5-II, ×600; all specimens from the phosphorite sample collected in upper courses of the Alekan River.

Comparison. Wide irregular thalluses differ these microfossils from the type species. In addition, smaller cells of *A. dactylographica* are always bacillary in shape and arranged in winding, but not radial series.

Material. A dozen of fragmented thalluses distinguished in two thin sections of calcareous-sandy phosphorite.

Locality. Identical to that of Holotype.

Reviewers V.N. Sergeev and M.A. Semikhatov

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The Lower Boundary of the Triassic and Its Correlation in Marine Sediments: Paper 1. Boundary Sections of the Tethys

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Abstract—In the Tethys region, the Triassic starts with the ammonite *Otoceras woodwardi* Zone accepted as basal in this system more than a hundred years ago. The boundary layers in Ciscaucasia, Iran, western Pakistan, Kashmir, central Himalayas, southern China, and Thailand are considered. The uppermost Permian deposits bearing *Paratirolites* or *Pseudotirolites* are well represented in the Paleotethys sections. In the Neotethys sections, the *Otoceras* genus often appears along with the *Ophiceras* genus and in association with conodonts *Hindeodus parvis*.

Key words: *uppermost Permian, lowermost Triassic, Tethyan sections, faunal assemblages, ammonite zones.*

Boundaries between the geologic systems as the most significant beacons in the earth's history have always attracted considerable attention. The boundary between the Permian and Triassic has been of particular concern as it coincides with the transition from the Paleozoic to Mesozoic, one of the most dramatic moments in the biosphere's evolution. In the last few years, the stratigraphic position and correlation of the boundary were at the center of theoretical debates that revealed a great variety of opinions and assessments. Some researchers, first of all specialists in ammonoids, advocate the traditional point of view as to draw the Permian–Triassic boundary at the base of the *Otoceras* Beds (table). Their opponents, adducing various arguments, propose that this boundary should be lowered and drawn at the base of the Dorashamian or, on the contrary, raised and drawn at the base of the *Ophiceras* Beds corresponding to the base of the upper Induan and even Olenekian. Especially active are geologists who insist on placing the lower boundary of the Triassic at the base of the conodont *Hindeodus parvus* Zone. Actually they try to replace the ammonite standard, which served as a foundation of the Triassic biostratigraphy, by the conodont scheme.

The problem of the Permian–Triassic boundary has gained momentum to such an extent that the International Subcommittee on the Triassic Stratigraphy organized in 1981 a special working group whose objective was to define the lower boundary of the Triassic and choose the global stratotype section and point for it.

The evolution of thoughts of this group is of interest. The inquiry carried out in 1984 showed that the majority of questioned scientists (16 out of 18) tend to recognize the lower boundary of the Triassic at the base of the *Otoceras* Beds. In contrast, 10 years later only two members of the group voted for this boundary, while 13 members considered the appearance of *H. parvus* as the initial event of the Triassic. The Meishan section in the Zhejiang Province, China, appeared to be the only one which the group recommended to be the global stratotype section and point for the Permian–Triassic boundary, although in August 1993, i.e., a year earlier, three other sections—Shangsi (Sichuan Province), Selong (southern Tibet) and Gurgyul Ravine (Kashmir)—were considered as possible candidates. The proposal by Professor Yin from China (Yin, 1985, 1993) to use *H. parvus* as the index species of the basal Triassic was supported by Kozur (1994) from Budapest, Hungary, as well as by Rachel and Richard Paull (Paull and Paull, 1994) from the University of Wisconsin–Milwaukee, USA. Some Russian researchers, who quite recently advocated the ammonite standard, for instance, Yu.D. Zakharov, appeared among supporters of the conodont boundary for the Triassic (e.g., in Kozur *et al.*, 1996).

In this connection, the history of identification of the Triassic lower boundary in marine deposits should be recalled. More than 100 years ago, Griesbach (1980), who carried out geological studies in central Himalayas, India, revealed beds with peculiar ammonites lying above the Permian Kuling Shale bearing *Productus* at the Shalshal Rock site near the Niti Pass (Painkhanda). He attributed the ammonites to new genera *Otoceras* and *Ophiceras* and named the deposits enclosing these ammonites the *Otoceras* Beds, considering them as the basal part of the Triassic.

At the beginning of the 20th century, a hot discussion on the age of the *Otoceras* Beds was initiated by Noetling (1900, 1904) in German geological journals. Mistaking *E. pisageceras dalailamae* Diener from the *Otoceras* Beds in the Himalayas for *E. wynnei* Waagen from the upper Salt Range Shale bearing *Productus*, he

made a conclusion that the enclosing rocks are of the same age and the *Otoceras* Beds belong to the Permian. Diener (1901, 1905) argued against his reasoning and used the presence of true Triassic ceratids and bivalves and the absence of the Permian brachiopods in the *Otoceras* Beds as the arguments in favor of their Triassic age. Spath (1934) shared his opinion later, and the Triassic age of the *Otoceras* Beds was not in doubt for the subsequent 50 years.

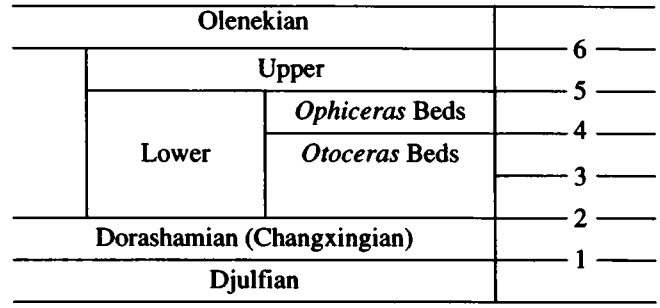
In the 1980s, a search for the global stratotype for the base of the Triassic has once again drawn attention to this boundary and caused new debates. Various alternatives that were suggested by some researchers have already been mentioned above (see table). The principal objections against the idea to place the boundary at the base of the *Otoceras* Beds can be summarized as follows: (1) otoceratids represent the latest descendants of the Permian araxoceratids, and ophiceratids are hard to distinguish from their Permian predecessors, xenodiscids; (2) localities with otoceratids are rare and spatially separated from localities with araxoceratids and *Paratirolites*; (3) judging from conodonts, the lower part of the *Otoceras* Beds corresponds to *Paratirolites* Beds and, thus, belongs to the Permian. Accordingly, it may be inferred that the traditional boundary of the Lower Triassic does not coincide with the beginning of the drastic fauna renewal, which happened later; it cannot serve, therefore, as the global demarcation line between the Permian and Triassic.

The *Otoceras* genus is really the last representative of the branch of ceratids originated in the Permian. It is likely that the genus is a direct descendant of the *Julfotoceras* genus occurring in the *Paratirolites* Beds of Iran (Bando, 1973). Both genera are connected as members of the continuous phylogenetic line *Julfoceras tarazi–Otoceras concavum–O. woodwardi*. This succession is largely hypothetical as no one succeeded in finding it anywhere in a single section.

Although ophiceratids from the *Otoceras* Beds resemble the Permian xenoceratids in the shape of their shells and lobe line structure, they differ radically from the latter in the ontogenesis (Shevyrev, 1990). If the suture line of xenodiscids follows the natural course of formation of secondary internal lobes, the ophiceratids show the line developing through the formation of umbilical lobes. Impressed by superficial similarity of lobe lines of xenodiscids and ophiceratids, neither Newell (1988, Fig. 7), nor paleontologists from China (Yang *et al.*, 1996, Figs. 3–5) have noticed these distinctions in the ontogenesis.

The spatial separateness of *Paratirolites (Pseudotirolites)* and *Otoceras* Beds presents certain problems in tracing the Triassic lower boundary. Deposits bearing *Paratirolites* (*Pseudotirolites*) are known in southern basins of the Paleotethys, whereas the *Otoceras* Beds are characteristic of the Boreal realm and basins of Neotethys (Fig. 1). No indisputable section, where the *Paratirolites (Pseudotirolites)* and *Otoceras* Beds occur in the direct contact, is known at present. Available attempts to correlate them, at least partially (Kozur, 1989, 1994), do not seem convincing, since the *Paratirolites* Beds bear an assemblage of typical Permian ammonites, which are missing from the *Otoceras* Beds, and the lobe line of the *Otoceras* is much more complex than that of predecessors from the *Paratirolites* Beds.

Alternative viewpoints on the position of the Permian–Triassic boundary



(1–6) Position of the boundary according to Sokratov, 1982 (1); 2. Nakazawa *et al.*, 1980, Dagys and Dagys, 1988, Tozer, 1988, 1994, Shevyrev, 1990 (2); Kozur *et al.*, 1996, Yin *et al.*, 1996 (3); Bhatt and Arora, 1984, Li *et al.*, 1989 (4); Newell, 1988 (5); Waterhouse, 1978 (6).

Assuming that the Permian–Triassic boundary is located between the mentioned beds, let us turn our attention to particular Tethyan sections, where the boundary deposits are represented most completely.

CISCAUCASIA

Ciscaucasia has attracted considerable interest of researchers since the beginning of the current century as a region of presumably continuous succession of marine deposits from the Permian to Triassic. The succession is most complete in the Dzhulfa Ravine of Nakhichevan, between the Dorasham-1 and Dorasham-2 stations, where it is represented by the boundary beds of the Akhura and Karabaglyar formations (Fig. 2). The sequence consists of the following beds:

(1) Brick–red limestone with thin interlayers of brown mudstone (4.0 m) makes up the member, which crowns the Akhura Formation. The member yielded tabulates, tetracorals, brachiopods, ammonites *Paratirolites kittli* Stoyanow, *P. waageni* (Stoyanow) and others, as well as conodonts *Clarkina subcarinata* (Sweet), *Hindeodus minutus* (Ellison) and forms transitional to *H. parvus* (Kozur et Pjatakova) (Kozur *et al.*, 1978; Kotlyar *et al.*, 1984; Zakharov, 1985; Zakharov and Rybalka, 1987).

(2–12) Brick–red mudstone with interlayers of brownish gray marl (2.0 m) represents basal deposits of the Karabaglyar Formation. The lowermost beds of the formation bear ammonites *Pleuronodoceras occidentale* Zakharov; Bed 5 (0.4 m above the base) includes *Xenodiscus jubilaearis* Zakharov; the uppermost beds

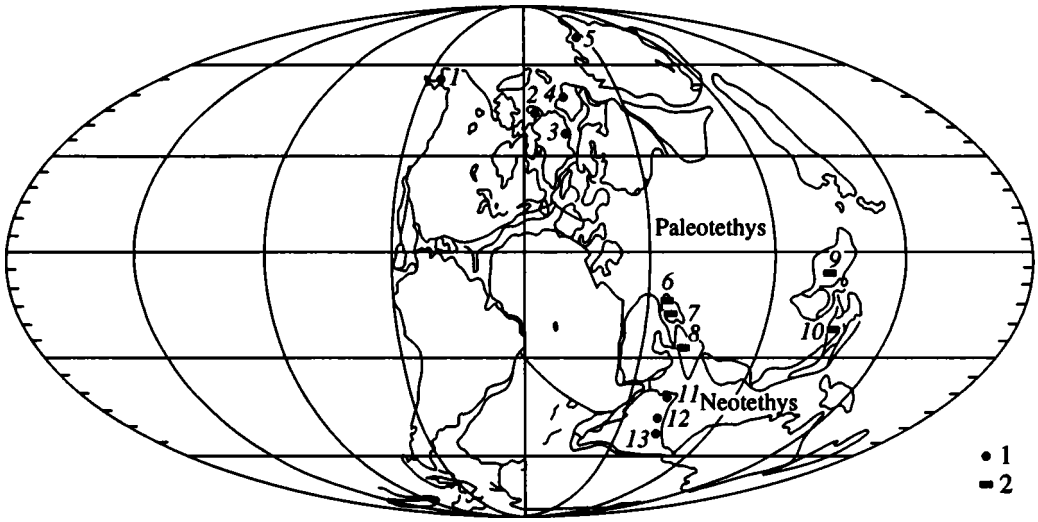


Fig. 1. Localities of the Permian–Triassic key ammonites in the palinspastic map (Scotese, McKerrow, 1990). (1) *Otoceras*; (2) *Paratirolites, Pseudotirolites*. Numbered regions: *1*, Alaska; *2*, Axel Heiburg and Ellesmere islands; *3*, eastern Greenland; *4*, Spitsbergen; *5*, eastern Verkhoynsk region; *6*, Ciscaucasia (Dzhulfa); *7*, northwestern Iran (Kuh-e-Ali Bashi); *8*, central Iran (Abadeh); *9*, southern China (Meishan, Shangsi); *10*, northern Thailand; *11*, Kashmir; *12*, southern Tibet (Selong); *13*, central Himalayas (Spiti, Painkhanda, Nepal)

contain conodonts *Hindeodus typicalis* (Sweet), *H. latidentatus* (Kozur, Mostler et Rahimi-Yazd), and *Clarkina changxingensis* (Wang et Wang).

(13–27) Yellowish green mudstone, greenish gray marl and limestone (2.0 m) represent a member bearing *Claraia intermedia* Bittner at different levels. Conodonts *H. parvus* are found in Bed 18 (0.65 m above the member base; *Ophiceras* sp. are identified in beds 21 and 24 (0.75 and 1.3 m above the base).

(28) Light-gray, massive limestone (0.17 m) of the bed yielded *H. parvus, Isarcicella isarcica* (Huckiede).

Zakharov (1988) included members 2–12 into the *Pleuronodoceras occidentale* Zone newly distinguished and assigned to the uppermost Permian. It should be pointed out that the index species of the zone is lacking marginal tubercles and differs from typical representatives of genus *Pleuronodoceras* from China. Its generic affinity still remains doubtful. Another ammonite species *X. jubilaearis* present in the zone indicates convincingly the Permian age of the unit. Ammonites are not encountered in the lower part of the *Claraia* Beds. At this level, conodonts *H. parvus* coexist with *Claraia*. Ammonites from genus *Ophiceras* and conodonts *I. isarcica* appear above this level.

IRAN

Sections similar to those from Ciscaucasia are located in northwestern Iran, in Kuh-e-Ali Bashi Mountains, 8 km south of the Dorasham-2 Station. One of them was studied by Stepanov and Iranian geologists in 1967–1968 (Stepanov *et al.*, 1969). Teichert and Kummel carried out the field work in this region in the summer of 1968 and studied four boundary sections (Teichert *et al.*, 1973). The red *Paratirolites* Limestone is 3.6 m thick here and bears ammonites *Paratirolites waageni, P. vediensis* Shevyrev, *Abichites mojsisovicsi* (Stoyanow), *Julfotoceras tarazi* Bando and others in association with brachiopods *Araxathyris araxensis minor* Grunt, *Spinimarginifera* sp., and tetracorals *Pleramplexus monimum* Iljina and *Plerophyllum* sp. The limestone is overlain without visible erosion marks by purple and violet marly shale (1.3 m), pinkish to yellowish platy limestone alternating with marly shale (2.0 m), and variegated limestone intercalated with marly shale interbeds (6.5 m). The last unit yielded bivalves *Claraia stachei* Bittner and *C. clarai* (Emmrich).

A thick section of the Upper Permian and Lower Triassic marine deposits was revealed in 1967 near Abadeh (central Iran), 1000 km southeast of Dzhulfa; it was described later by Taraz from the Geological Survey of Iran (Taraz, 1969, 1971). The joint Iranian–Japanese research group studied this region in detail in 1972–1975 (*Iranian–Japanese Research Group*, 1981).

The Upper Permian shaly–carbonate sequence ends here with brownish red limestone of the Hambast Formation. The limestone Member 7 is 17 m thick (Fig. 3). The member is subdivided into the *Shevyrevites* (5 m), *Shevyrevites–Paratirolites* (8 m), and *Paratirolites* (4 m) ammonite zones. The second and third zones include representatives of the genus *Julfotoceras*, the direct predecessor of genus *Otoceras* (Bando, 1973, 1979). Two conodont zones have been established in the same member: those of *Condonella orientalis* and *Anchignathodus (=Hindeodus) julfensis*. The boundary

between them is within the *Shevyrevites–Paratirolites* Zone of the ammonite scale.

The *Paratirolites* Beds are overlain by a sequence of the Lower Triassic limestone and shale of the Elikah Formation. The lower part of the sequence (Unit a) is composed of yellowish to grayish limestone and shale 90 m thick. Its base is made up of greenish to yellowish gray, locally pinkish, barren shale (0.1–0.3 m) with single stromatolites. They are replaced upsection by stromatolitic limestone (1.3–1.8 m) also bearing *Hypophiceras* sp., *Claraia* aff. *dalpiazi* Leonardi, *Hindeodus typicalis* (Sweet), and *H. parvus* (Kozur et Pjatakova). A greater part of the unit is composed of dark or light gray laminated limestone and greenish shale. *"Vishnuites" pralambha* Diener and *Ophiceras dubium* Spath were collected from a horizon 9.5 m above the base of the Triassic sequence (Band, 1979). Bivalves and conodonts are encountered at various levels. The latter discriminate the *parvus* and *isarcica* zones.

WESTERN PAKISTAN

Many litho- and biostratigraphic features of boundary deposits in this region remained unclear for a long time. The deposits were studied by Kummel and Teichert in the 1960s (Kummel and Teichert, 1966, 1970; Kummel, 1970), who focused their attention on the Upper Permian Chhidru Formation and Kathwai Member of the Triassic Mianwali Formation. The joint research group from Pakistan and Japan investigated the region in 1975–1979 (*Pakistani–Japanese Research Group*, 1985). From the uppermost parts of the Chhidru Formation, the group has collected samples with the brachiopod assemblage of *Oldhamina decepiens* Koninck, *Richthofenia lawrenciana* Koninck, *Enteletes socialis* Reed, and *Whitspakia acutangula* (Waagen). The beds with the assemblage have been attributed to the *socialis* Zone of the Djulfian Stage (Fig. 4).

The research group from Pakistan and Japan subdivided the Kathwai Member into three parts. The lower part is composed of dolomite, dolomitized limestone, and calcareous sandstone intercalated with coquina interlayers consisting of fragmented brachiopod shells, echinoderms, and bryozoans. The thickeness varies from 0 to 1.6 m. Found in this part are brachiopods *Enteletes* sp, *Orthothetina* cf. *arakeljani* Sokolskaya, *Ombonia* sp., *Lyttonia* sp., *Crurithyris ? extinis* Grant, and others associated with foraminifers (*Globivalvulina* sp., *Nodosaria* sp., *Pachyphloia* sp., *Reichelina* sp.), bryozoans (*Dyscritella* ? sp., *Fistulipora* sp.), phenestellids, and very rare conodonts *H. minutus* suggesting that these beds correspond to the *minutus* Zone.

The middle part of the Kathwai Member is 0.8–1.5 m thick and composed of reddish brown or orange-colored, massive to foliated dolomite with interlayers of pale green mudstone. This part yielded smaller foraminifers *Hemigordius* sp., *Staffella* sp., *Pachyphloia* sp., *Paradagmarita* sp., and others associated with bivalves (*Eumorphotis waageni* (Hauer) and *Pernopecten* sp.) and conodonts *Hindeodus minutus, H. parvus, Isarcicella isarcica*, and *Clarkina carinata*, all characteristic of the *parvus–isarcica* Zone. This zone was later subdivided into the *parvus–minutus* and *isarcica–parvus* subzones (Nakazawa, 1993). Specimens of *Hypophiceras himalayanum* (Griesbach) seem to be found by Kummel and Teichert in this zone.

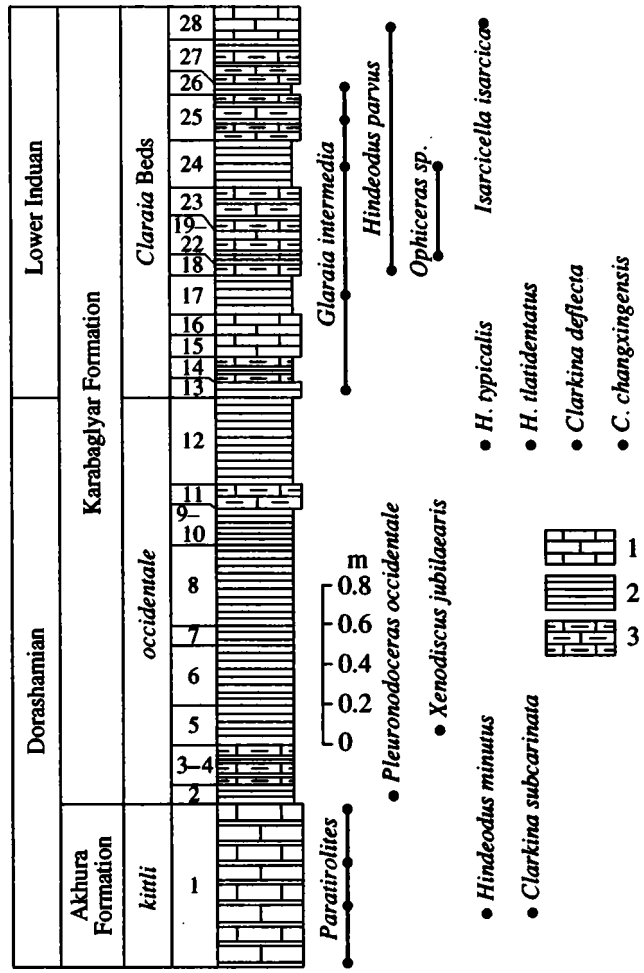


Fig. 2. Section of the Permian–Triassic boundary deposits in the Dzhulfa Ravine, near the Dorasham-2 Station (Zakharov, 1985; Zakharov and Rybalka, 1987; Zakharov, 1988): (1) limestone; (2) mudstone (shale); (3) marl.

The upper part of the Kathwai Member is variable in its lithologic composition and thickness. In the Narmia section (Surghar Range), it consists exclusively of foliated limestone 0.7 m thick. In the Chhidru section, where its thickness is 3.6 m, the unit is composed of alternating dolomite and shale in the lower part and of thin-bedded sandstone and shale intercalated with interlayers of limestone in the upper part. In other sections, this part, 1.8–2.3 m thick, is represented by lower dolomite and upper limestone beds. The conodont spe-

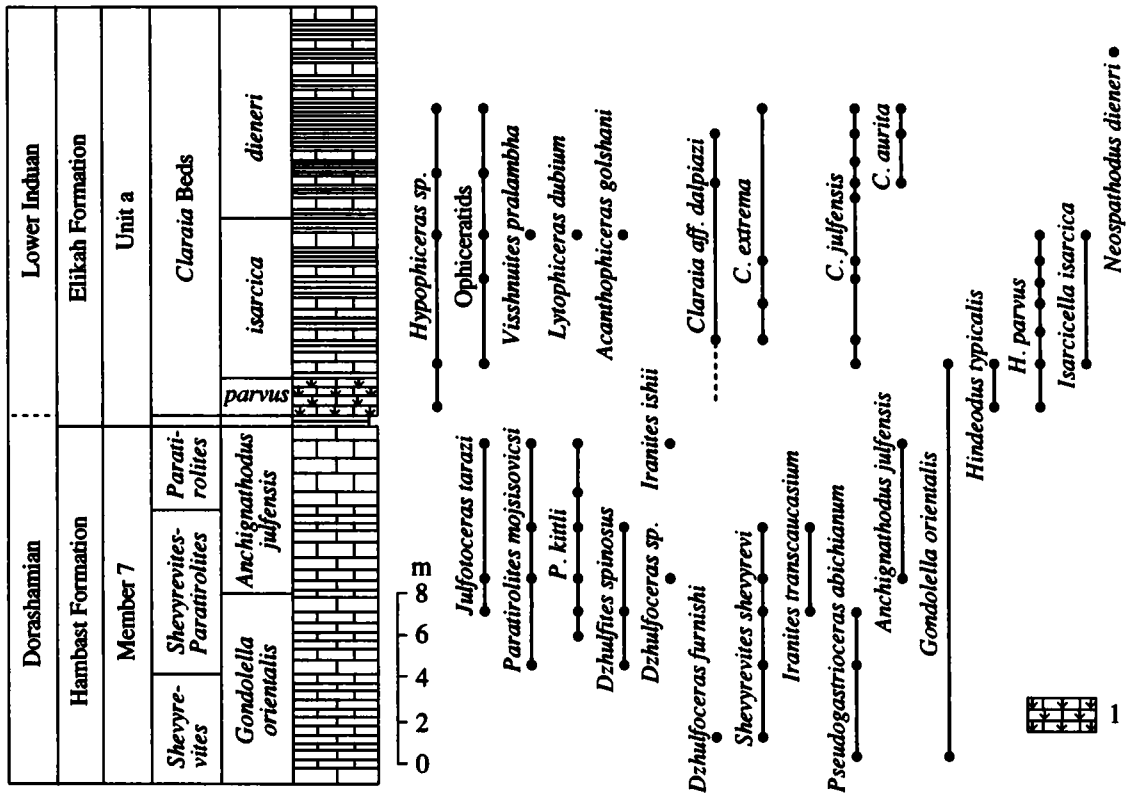


Fig. 3. Section of the Permian–Triassic boundary deposits in central Iran near Abadeh (Taraz, 1969, 1971; Bando, 1979; *Iranian–Japanese Research Group*, 1981): (1) stromatolithic limestone (other symbols as in Fig. 2).

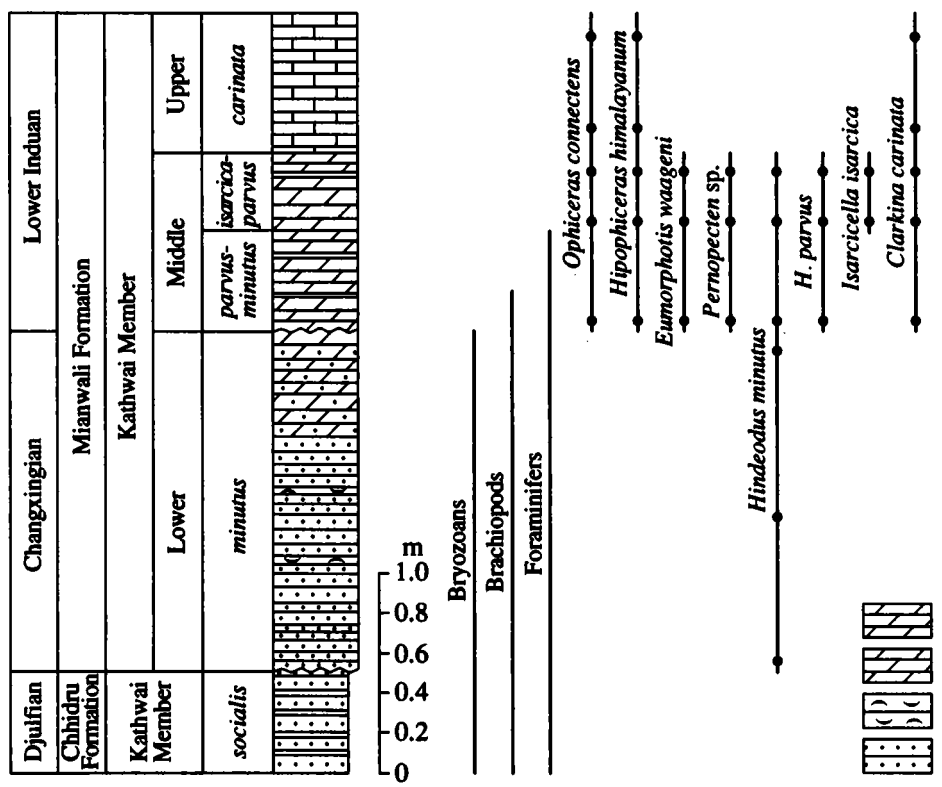


Fig. 4. Section of the Permian–Triassic boundary deposits in the Salt Range (*Pakistani–Japanese Research Group*, 1985): (1) dolomite; (2) sandy dolomite; (3) coquina; (4) sandstone (other symbols as in Fig. 2).

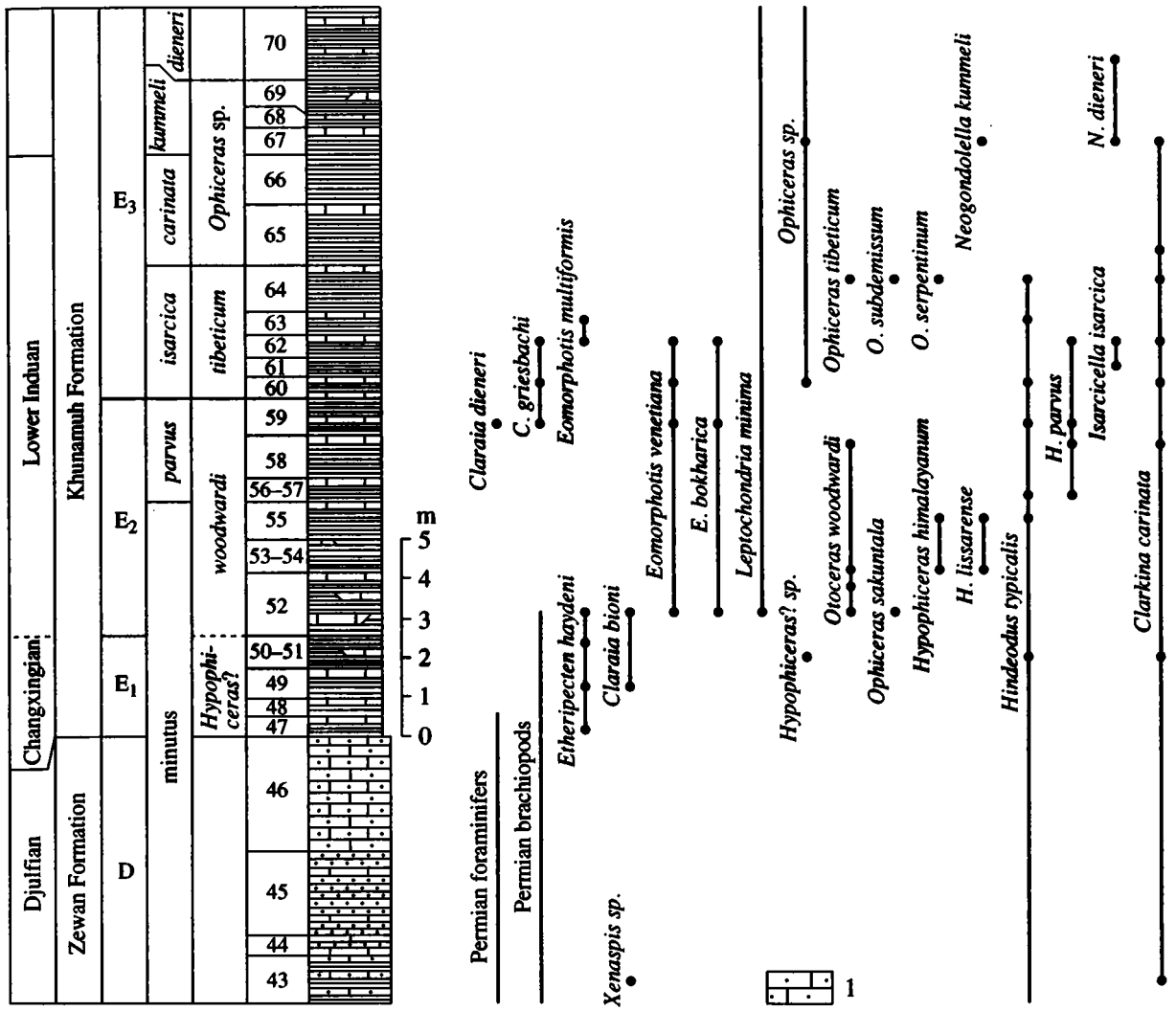


Fig. 5. Section of the Permian–Triassic boundary deposits in Kashmir, the Gurgyul Ravine (Nakazawa*et al.*, 1975; Nakazawa, 1981, 1993; Kapoor, 1996): (1) sandy limestone (other symbols as in Fig. 2).

cies *Clarkina carinata* is known from the section under consideration. *Ophiceras connectens* Schindewolf and *Ophiceras* sp. are found in the middle and upper parts of the section.

KASHMIR

An outcrop in the Gurgyul Ravine located east of Srinagar is recognized as one of the plausible candidates of the global stratotype section and point (GSSP) for the Triassic lower boundary. The study of the outcrop began as early as the second half of the past century, continued shortly after, and was resumed only in 1969 by Japanese and Indian geologists, who studied it comprehensively during several years (Nakazawa *et al.*, 1975; Nakazawa, 1981, 1993; Nakazawa and Kapoor, 1981; Kapoor, 1992, 1996).

In the ravine, the Permian sequence ends with the Zewan Formation subdivided into four lithologic members. The upper one (Member D, beds 32–46) is made up of clayey sandstone, sandy limestone, and shale beds 18.2 m thick (Fig. 5). The member is crowned by a bed of brown sandy limestone 2.9 m thick. Some levels show mass accumulations of bivalves *Permophorus* cf. *subovalis* (Waagen), *Etheripecten haydeni* Nakazawa, *Cyrtorostra* aff. *lunwalensis* (Reed), and others coexisting with gastropods *Bellerophon branfordianus* Waagen and *Retispira ornatissima* Waagen, and with brachiopods *Linoproductus* cf. *lineatus* (Waagen) and *Waagenoconcha* sp. Foraminifers; bryozoans are rare. Conodonts *Clarkina carinata* and *Ellisonia triassica* are identified in the middle part of the member (Bed 43). Ammonite of *Xenaspis* sp. is found in the same bed.

Since the index fossils are missing, the age of the upper member of the Zewan Formation remains undetermined. It is referred to either the uppermost Djul-

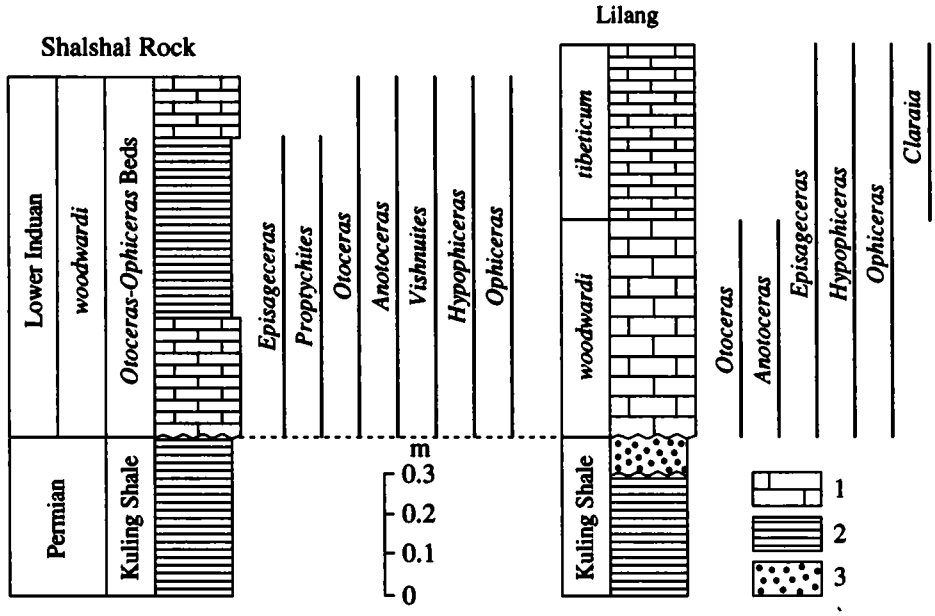


Fig. 6. Section of the Permian–Triassic deposits in central Himalayas (Diener, 1912; Kummel, 1972; Bhatt and Joshi, 1981): (1) limestone; (2) shale; (3) conglomerate.

fian–lowermost Dorashamian (Nakazawa *et al.*, 1975, Plate 13) or the Changxingian (Nakazawa, 1993, Fig. 10; Kapoor, 1996, Plate 8.1).

The overlying Khunamuh Formation is made up of alternating limestone and shale. It is subdivided into several members of different lithology. The lower Member E (beds 47–70) is composed of shale with limestone interlayers; its total thickness is 18.6 m. It is subdivided into three submembers according to the shale type.

The lowermost Member E_1 (beds 47–51) is represented by dark-gray and black shale with thin intermittent limestone layers 2.6 m thick. The shale yielded the Permian smaller brachiopods *Linoproductus* cf. *lineatus* (Waagen), *Marginifera himalayensis* Diener, and others along with bivalves *Etheripecten haydeni* Nakazawa, *E.* cf. *hiemalis* (Salter), *Cyrtorostra* aff. *lunwalensis* (Reed), and *Claraia bioni* Nakazawa; rare foraminifers *Nodosinella longissima* M. Maclay and *Lunucammina* sp.; few poorly preserved conodonts *Clarkina carinata*; and *Hindeodus typicalis*. Among undeterminable ammonites of the member (xenodiscids or ophiceratids), there are problematic *Hypophiceras* forms. This part of the section is identified as the *Hypophiceras* ? Zone.

The middle Member E_2 (beds 52–59; 6.1 m thick) is composed of black and dark-gray platy shale alternating with limestone interbeds, which bear the following fossils: bivalves *Eumorphotis venetiana* (Haiser), *Leptochondria minima* (Kiparisova), *Claraia dieneri* Nakazawa, *C.* cf. *griesbachi* (Bittner), and *C.* aff. *extrema* Spath; ammonites *Otoceras woodwardi* Griesbach, *Ophiceras sakuntala* Diener, *Hypophiceras himalayanum* (Griesbach), and *H. lissarense* (Diener); bellerophontid *Waithia hisakatsui* Murata; conodonts *Hindeodus typicalis*, *H. parvus*, and *Clarkina carinata*. This part of the section is attributed as a whole to the ammonite *woodwardi* Zone, but its upper part (beds 56–59) is considered as corresponding to the *parvus* Zone of the conodont zonation.

The upper Member E_3 (beds 60–70; 9.9 m thick) is made up of black shale alternating with limestone interbeds. The shale is less platy than that of Member E_2. The lower beds yielded the diverse fossils: ammonites *Ophiceras tibeticum* Griesbach, *O. subdemissum* (Spath), and *O. serpentinum* Diener; bivalves *Claraia* aff. *extrema* Spath, *C.* cf. *griesbachi* (Bittner), *Eumorphotis venetiana* (Hauer), and *E. multiformis* (Bittner); conodonts *H. parvus*, *H. typicalis*, *Clarkina carinata*, *Isarcicella isarcica, and Neospathodus kummeli*; foraminifers *Glomospirella* cf. *shengi* and rotaliids. The upper beds bear *Ophiceras* sp., *Claraia concentrica* Yabe, and *Leptochondria minima* Kiparisova. Distinguished in this part of the section are the *tibeticum* (beds 60–64) and *Ophiceras* (beds 65–69) ammonite zones, as well as *isarcica* (beds 60–64), *carinata* (beds 65–66), *kummeli* (beds 67–69), and *dieneri* (Bed 70) zones of the conodont scale.

CENTRAL HIMALAYAS OF INDIA

The *Otoceras*, or to be more exact, the *Otoceras–Ophiceras* Beds of the Shalshal Rock (Painkhanda) representing the stratotype of the *woodwardi* Zone (Fig. 6), are subdivided into the following units: black shale bed (0.3 m) with *Otoceras woodwardi* Griesbach, *Ophiceras tibeticus* Griesbach, and *Episageceras*

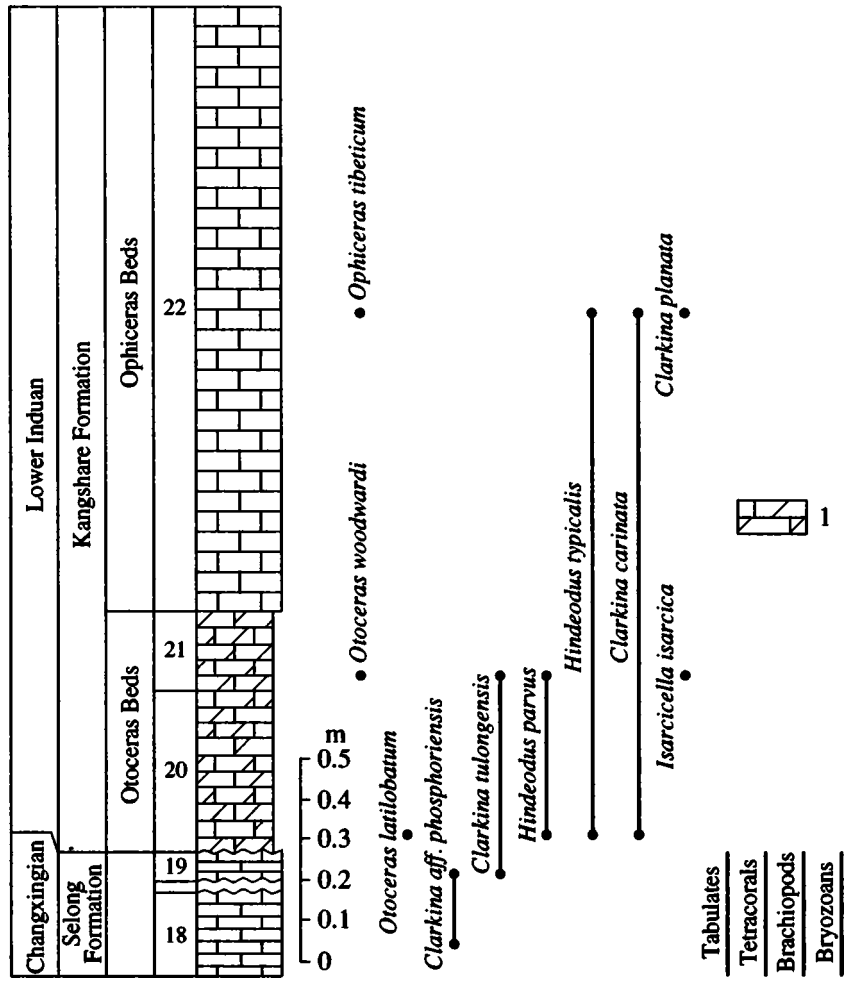


Fig. 7. Section of the Permian–Triassic boundary deposits in southern Tibet near the Selong Settlement (Jin *et al.*, 1996): (1) dolomitized limestone (other symbols as in Figs. 2 and 4).

dalailamae Diener; shale bed (0.45 m) with *E. dalailamae* and *Pseudoproptychites scheibleri* (Diener); and black limestone bed (0.15 m) with *O. woodwardi*, *Ophiceras sakuntala* Diener, and *O. tibeticum* Griesbach (Diener, 1912; Kummel, 1972). Genera *Hypophiceras*, *Vishnuites*, and *Anotoceras*, all typical of the Triassic and unknown in the Permian, are dispersed throughout the sequence of these beds corresponding to the *woodwardi* Zone. In the Lilang (Spiti) section, the *Otoceras–Ophiceras* beds are similarly thin and composed of dense massive limestone (0.54 m) with *O. woodwardi*, *Hypophiceras himalayanum* (Griesbach), *Anotoceras nala* Diener, *Ophiceras sakuntala* Diener, and *Episageceras dalailamae*, and also of platy limestone (0.44 m) with *Ophiceras tibeticum*, *Hypophiceras himalayanum*, *Episageceras dalailamae*, and *Claraia* shells (Kummel, 1972; Bhatt and Joshi, 1981). The beds can be subdivided here into the *woodwardi* (massive limestone with *Otoceras* and *Ophiceras*) and *tibeticum* (platy limestone with *Ophiceras* but without *Otoceras*) zones. In both sections, the *Otoceras* Beds unconformably overlie the Permian Kuling Shale.

SELONG

The Selong section is located in southern Tibet near the Selong Settlement (in the neighbourhood of Karru-Ochen Lake), 4566 m above the sea level. Revealed by the Chinese research group in the mid-1970s, the section was suggested as a potential GSSP of the Permian–Triassic boundary (Wang *et al.*, 1988; Jin *et al.*, 1996).

The boundary deposits belong here to the Selong and Kangshare formations (Fig. 7). The Selong Formation is composed of calcareous sandstone, mudstone, and crinoid limestone beds about 100 m thick. It is crowned with the coral bed (Bed 18, 0.2 m thick) of foliated limestone with abundant tabulates and tetracorals, fragments of crinoids, brachiopods, bivalves and bryozoans. Brachiopods characterize the *Chonetella nasuta* (Waagen) assemblage. Conodonts *Clarkina* aff. *phosphoriensis* are found here as well. The caliche or

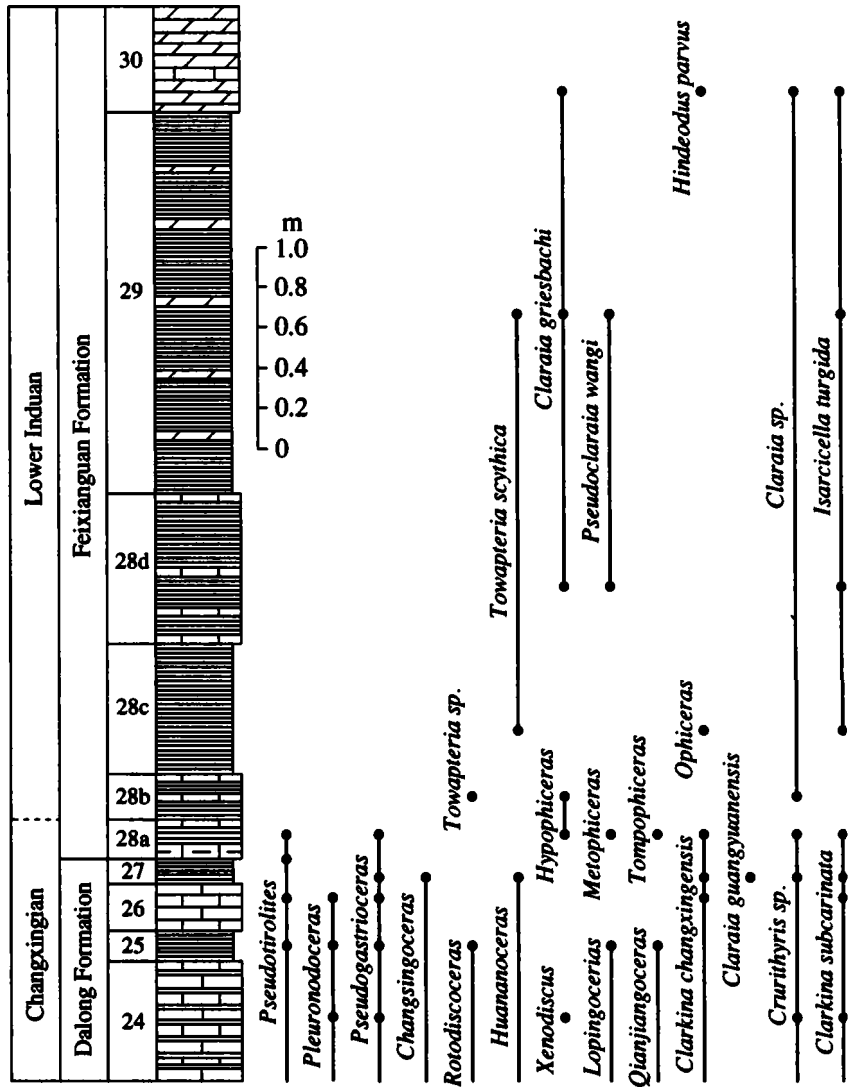


Fig. 8. Section of the Permian–Triassic boundary deposits in Sichuan Province, Shangsi (Lai *et al.*, 1996; symbols as in Figs. 2 and 4).

clay bed (up to 3.5 cm thick) between the Selong and Kangshare formations is replaced up-section by the *Waagenites* Limestone (Bed 19, 0.07 m thick) of brownish yellow color, which yielded the diverse fossils: brachiopods *Waagenites* cf. *barusiensis* (Davidson), *Paracrurithyris pygmaea* Liao, *Phricodothyris* sp., *Araxathyris* sp., and *Martinia* sp. (the *W. barusiensis–P. pygmaea* assemblage); conodonts *Clarkina* ex gr. *phosphoriensis* and *Vjalovognathus* sp.; fragments of bryozoans, gastropods, and ostracodes. The bed is attributed either to the Selong Formation (Wang *et al.*, 1988) or to Kangshare Formation (Jin *et al.*, 1996).

With the hiatus, in the opinion of some researchers, or conformably, in the view of others, the *Waagenites* Limestone overlies the brownish yellow dolomitized limestone with ammonites *Otoceras latilobatum* Wang and conodonts *Hindeodus parvus, H. typicalis*, and *Clarkina carinata.* The overlying gray to yellow dolomitized limestone (Bed 21, 0.2 m thick) yielded *Otoceras woodwardi* and conodonts *H. parvus, Clarkina tulongensis, C. taylori,* and *Isarcicella isarcica* (Jin *et al.*, 1996). The last unit is overlain conformably or with a break by yellow to gray bioclastic limestone (Bed 22, 1.6 m thick) bearing *Ophiceras tibeticum, Clarkina carinata, C. planata,* and *Hindeodus typicalis.*

SHANGSI

The section, named sometimes as Changjianggou locality, is situated near Guangyuan City, in the north of Sichuan Province. Revealed in the 1960s, it was subjected to comprehensive study in the 1980s (Yang *et al.*, 1987; Li *et al.*, 1989; Lai *et al.*, 1996). In 1993, the working group suggested the section, along with others, as a plausible GSSP of the Triassic lower boundary.

The boundary deposits occur here between the Dalong and Feixianguan formations (Fig. 8). The uppermost parts of the Dalong Formation consists of the following beds:

(24) dark-gray siliceous limestone with interbeds of siliceous shale (0.60 m); the bed yielded ammonites *Pleuronodoceras tenuicostatum* Chao, Liang et Zheng, *Pseudogastrioceras* aff. *guizhouense* Chao, Liang et Zheng, and *Xenodiscus* sp. in association with brachiopods *Crurithyris pusilla* Chen and conodonts *Clarkina subcarinata* (Sweet);

(25) yellow-gray to greenish white illite–montmorillonite clay (0.15 m) with *Pleuronodoceras* sp., *Rotodiscoceras* sp., *Pseudotirolites* cf. *acutus* Chao et Liang, *Pseudogastrioceras* sp., *Qianjiangoceras* sp., and *Lopingoceras* sp;

(26) dark-gray foliated siliceous micrite (0.25 m) with *Pseudogastrioceras* sp., *Pseudotirolites disconnectus* Chao, Liang et Zheng, *Pleuronodoceras* sp., *Clarkina subcarinata* (Sweet), *C. changxingensis* (Wang et Wang), and *C. guangyanensis* (Dai et Tien);

(27a) grayish black montmorillonite and illite clay (0.05 m) with *Clarkina changxingensis, C. subcarinata, C. tulongensis* (Wang et Wang), *Crurithyris changjiangouensis* Zhan, *C. pygmaea* Liao, and *Pseudogastrioceras* sp.;

(27b) grayish white illite and montmorillonite clay with volcanogenic admixture (0.06 m);

(27c) grayish black illite and montmorillonite clay (0.04 m) with *Claraia guangyuanensis* Li, *Huananoceras* sp., *Pseudotirolites asiaticus* (Jaeckel), *Changxingoceras* sp., *Clarkina subcarinata*, and *C. changxingensis.*

The Dalong Formation is conformably overlain by the following beds of the Feixianguan Formation:

(28a) yellowish green foliated illite and montmorillonite clay with interbeds of calcareous shale and marl (0.16 m); the bed bears *Hypophiceras* sp., *Metophiceras* sp., *Tompophiceras* sp., *Pseudotirolites* sp., *Pseudogastrioceras* sp., *Crurithyris* sp., *Clarkina changxingensis, C. subcarinata,* and *Hindeodella* sp.;

(28b) light-yellow or yellowish green silty and montmorillonite clay with interbeds of foliated micrite (0.24 m); fossils are represented by *Claraia* sp., *Towapteria* sp., and *Hypophiceras* sp.;

(28c) yellowish green silty illite clay (0.64 m) with *Isarcicella turgida* (Kozur, Mostler et Rahimi-Yazd), *Claraia* sp., *Towapteria scyhica* (Wirth), and *Ophiceras* cf. *sinense* Tien;

(28d) grayish green silty illite clay alternating with calcareous shale (0.74 m); fossils of the bed are *Bakevellia* sp., *Claraia hubeiensis* Chen, *C. yunnanensis* (Jin et Hsu), *Pseudoclaraia wangi* (Patte), and *Isarcicella turgida*;

(29) yellowish gray foliated illite and montmorillonite clay with marl interbeds (1.90 m); the bed yielded *Isarcicella turgida, Hindeodus minutus* (Ellison), *Neohindeodella triassica* (Müller), *Claraia griesbachi* (Bittner), *Pseudoclaraia wangi*, and *Towapteria scythica*;

(30) yellowish gray foliated marl with interbeds of gray clayey micrite (1.8 m) and remains of *Hindeodus parvus, Isarcicella turgida, Xaniognathodus elongatus* Sweet, and *Claraia griesbachi.*

MEISHAN

The Meishan section is exposed in seven limestone quarries situated in the southern slope of the Mt. Meishan (Changxingian District, Zhejiang Province). Sections D (Baoqing, the Changxingian stratotype) and Z (Zhongxing Dadui) are spaced 0.5 km from one another and studied more comprehensively than others. There is no general agreement among Chinese geologists as to which of these sections should be chosen as the GSSP of the lower boundary of the Triassic. Some of them give preference to the first quarry (Yang *et al.*, 1987; Yin *et al.*, 1994) while the others favour the second (Sheng *et al.*, 1984; Wang, 1995). Section D is more complete and studied in detail (Yin *et al.*, 1996). It is under the state of protection now.

The Permian Changxingian Formation terminates with the Meishan Member composed of gray biomicritic limestone 11 m thick. Its top is made up of Bed 24 (Fig. 9) into five units:

(24a) dark-gray laminated limestone (0.10 m) with a thin upper clay layer and conodonts *Clarkina carinata* (Clark) and *C. changxingensis* (Wang et Wang);

(24b) dark-gray dolomitized limestone (0.11 m) with a thin clay layer at the top and conodonts *Clarkina deflecta* (Wang et Wang) and *C. changxingensis*;

(24c) dark-gray dolomitized limestone (0.17 m) with *Enantiognathus ziegleri* (Diebel), *Hibbardella* sp., and *Xaniognathus elongatus* Sweet.

(24d) dark-gray limestone with interlayers of calcareous mudstone (0.23 m) and conodonts *Enantiognathus ziegleri, Clarkina carinata, C. deflecta,* and *C. changxingensis*; foraminifers of the unit are *Geinitzina caucasica* M. Maclay, *Nodosaria netchajevi* Lipina, *Palaeofusulina* cf. *sinensis* Sheng, and *Pachyphloia lanceolata* M. Maclay; ammonites *Pleuronodoceras mirificum* Zhao, Liang et Zheng, and *Pseudogastrioceras* sp. also occur here;

(24e) Gray limestone (0.20 m) with fusulinids *Palaeofusulina* sp., ammonites *Rotodiscoceras* sp., conodonts *Clarkina changxingensis, C. deflecta, C. carinata,* and *C. subcarinata*; coexisting brachiopods are *Crurithyris flabelliformis* Liao, *Neowellerella pseudoutah* (Huang), and *Wellerella delicatula* Dunbar et Condra.

The Yinkeng Formation (also known under names Qinglong or Chinglung) rests above the last unit. Basal in the formation is the composite Bed 1 (lower transitional bed) consisting of two units:

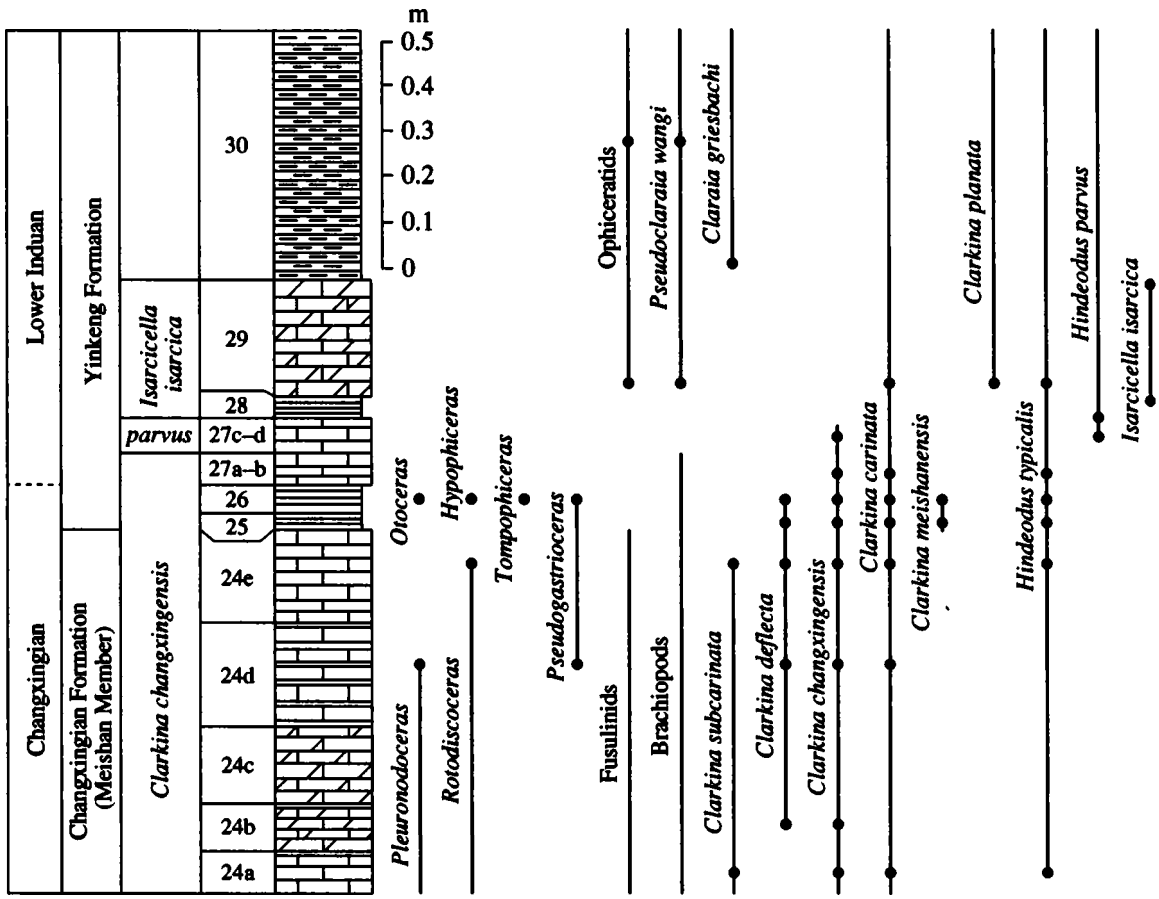


Fig. 9. Section of the Permian–Triassic boundary deposits in Zhejiang Province, Meishan (Yin *et al.*, 1996; symbols as in Figs. 2 and 7).

(25) illite–montmorillonite (0.04 m) clay known as White Clay, bluish gray, turning yellowish white at the weathered surface and bearing conodonts *Clarkina changxingensis, C. deflecta, C.* cf. *carinata, C. orientalis* (Barskov et Kovaleva), *C. subcarinata, Hindeodus typicalis,* and *H. latidentatus* in association with foraminifers *Bradyina* sp., *Glomospira* sp., *Globivalvulina* sp., *Nodosaria* sp., and others;

(26) dark-gray montmorillonite–illite clay (0.06 m) known as Black Shale and containing brachiopods *Cathaysia chonetoides* (Chao), *Crurithyris flabelliformis* Liao, *Neochonetes convexa* (Liao), *Paryphella orbicularis* (Liao), and *Waagenites wongiana* (Chao), which coexist with conodonts *Clarkina changxingensis, C. deflecta, C. carinata, C. meishanensis* Zhang, Lai *et al.* Fauna from this level of section Z is represente by several groups of fossils: ammonites *Otoceras* ? sp., *Hypophiceras* cf. *martini* Trümpy, *H. changxingense* Wang, *Metophiceras* sp., *Pseudogastripceras* sp., and *Tompophiceras* sp.; bivalves *Peribositra baoqingensis* Chen; brachiopods *Araxathyris minuta* Grunt, *Paracrurithyris pigmaea* (Liao), *Paryphella sulcatifera* (Liao), and *Waagenites barusiensis* (Davidson); conodonts *Hindeodus latidentatus.*

Higher in the section, there are the following beds:

(27a–b) light-gray silty limestone (0.08 m) with conodonts *Hindeodus typicalis, H. latidentatus, Merrillina longidentata* Kozur, and *Stepanovites* sp.;

(27c–d) similar limestone (0.08 m) with *Hindeodus parvus, H. latidentatus, Isarcicella turgida, Clarkina changxingensis,* and *C. carinata* in Bed 27c and with *H. parvus, H. typicalis, Paryphella triquetra* Liao, *P. orbicularis* (Liao), *Waagenites* sp., *Paracrurithyris* sp., and *Fusichonetes pigmaea* Liao in Bed 27d;

(28) grayish yellow illite–montmorillonite clay (0.04 m) with *H. parvus, Isarcicella turgida, I. isarcica,* and *Clarkina carinata*;

(29) gray dolomitized limestone (0.26 m) with ophiceratids; bivalves *Pseudoclaraia wangi* (Patte); brachiopods *Paryphella orbicularis*; and conodonts *Clarkina planulata* (Clark), *C. carinata, H. typicalis,* and *Isarcicella turgida*;

(30) bluish gray marl (0.55 m) with ophiceratids and bivalves *Pseudoclaraia wangi, Claraia concentrica* (Yabe), and *C. griesbachi* (Bittner).

NORTHERN THAILAND

An intriguing section of the Permian–Triassic boundary deposits was described at the Doi Pha Phlung locality near Lampang (Ishibashi and Chonglakmani, 1990; Ishibashi *et al.*, 1994).

The Permian Huai Thak Formation 1100 m thick is composed here of terrigenous and carbonate rocks with interbeds of chert and tuff. The upper part of the formation, 100 m below the boundary, consists of limestone beds with *Palaeofusulina sinensis* Sheng, *Reichelina* aff. *changxingensis* Sheng et Zhang and other upper Changxingian foraminifers. The formation is crowned with black shale bed bearing *Paratirolites nakornsrii* Ishibashi et Chonglakmani, *Tapashanites yaowalakae* Ishibashi, Nakornsri et Nagai, *Pseudogastrioceras* aff. *szechuanense* Chao et Liang, and *Xenodiscus* ? sp., i.e. the typical ammonite assemblage of the uppermost Changxingian. It is found 10 m below the Permian–Triassic boundary which coincides here with a minor fault.

The Triassic Phra That Formation is made up of black shale, sandstone, and limestone. *Ophiceras sakuntala* Diener, *O.* cf. *subkyokticum* Spath, and *O. chamunda* Diener were collected at the level of 1 m above the boundary, and *Claraia* cf. *ovata* (Schauroth), *C. stachei* (Bittner), *C. intermedia* (Bittner) and other bivalves were sampled at the 20-m level above it.

The sections described can be divided into two groups. The first group includes those sections where the uppermost Permian horizons are well developed and characterized by paratirolitids and pseudotirolitids. At the same time, the *Otoceras* Beds corresponding to the lowermost Triassic are missing here. In these sections, the Triassic sequence begins with the *Claraia* or *Ophiceras* Beds, i.e., with deposits of a higher level. The group incorporates sections of Ciscaucasia, Iran, southern China (Shangsi, Meishan), and Thailand. The second group includes sections lacking the *Paratirolites* Beds, and the basal Triassic is represented here by the *Otoceras* Beds. In this case, *Otoceras* forms appear usually along with *Ophiceras* species and *Hindeodus parvus*, locally even earlier. This situation is characteristic of sections in Kashmir (Gurgyul Ravine), central Himalayas (Shalshal Rock, Lilang, Nepal), and southern Tibet (Selong). It is of interest that the sections of the first group are confined to southern basins of the Paleotethys, while those of the second group are typical of the Neotethys region (see Fig. 1).

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Stratigraphy and the Upper Jurassic–Lower Cretaceous Radiolarians from the Carbonate–Siliceous Deposits, Ankara Region, Turkey

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Abstract—In the Ankara region, the Jurassic–Cretaceous Alacaatli key section spans the stratigraphic interval from the Oxfordian to the Valanginian (but not all the three Jurassic Series, as it was believed earlier). Abundant and diverse radiolarian assemblages are used for the detailed subdivision of these deposits and their correlation with the sections in the West Mediterranean region. Three units were distinguished and dated by radiolarians: (1) the middle Oxfordian–lower Tithonian (cherts and limestones, including the "ammonitico rosso" facies of the latter), 35–40 m; (2) the upper Kimmeridgian–Lower Berriasian (limestones and marls), 33 m; (3) the upper Tithonian–upper Valanginian (micritic limestones and calcarenites), 59 m.

Key words: *radiolarians, biostratigraphy, Upper Jurassic, Lower Cretaceous, foraminifers, Turkey, the Mediterranean area.*

The Jurassic and Lower Cretaceous sedimentary rocks are characteristic of the central regions of Turkey located to the north, southeast, and southwest of Ankara (Fig. 1a). They represent a thick transgressive sequence overlying, with a hiatus and basal conglomerates, the low-grade metamophic terrigenous–carbonate beds of the Permian–Triassic Karakaya Formation. In certain cases, the relationships between the Jurassic sediments and the Karakaya Formation are tectonic. The lower part of the section (Lias–Dogger) is mainly composed of terrigenous sediments with interbeds and lenses of carbonate rocks, whereas the upper part (Malm and Lower Cretaceous) consists of carbonates with siliceous interbeds. The Upper Jurassic part of the section yielded abundant radiolarians inadequately studied so far.

In the vicinity of the Alacaatli Village (15 km southeast of Ankara; Fig. 1b), a section was described earlier, the lower part of which is composed of calcareous cherts and siliceous limestones with interbeds and lenses of red limestones of the "ammonitico rosso" facies. The middle of the section shows a succession of yellowish beds of clay limestones and marls, and the upper part consists of white micritic limestones (Batman *et al.*, 1977). It was assumed that these beds overlie the Permian–Triassic Karakaya Formation and span three series of the Jurassic system; the lower unit was referred to the Lias, the middle to the Dogger, and the upper to the Malm. In the cited publication, the paleontological remains, however, are insufficiently represented and sometimes unusable for the age determination of enclosing rocks. For example, the Early Jurassic age of the lower part of the Alacaatli section is inferred only from occurrence of ammonites *Perisphinctes* sp., and *Harpoceras* sp.. Moreover, the exact position of these finds has not been indicated (Batman *et al.*, 1977). Likewise, no one studied systematically the ammonoids and belemnites found at different levels of the Alacaatli section.

The entire Alacaatli section includes abundant radiolarians recovered from the siliceous and micritic limestone beds. In recent years, data on radiolarians were actively used to elucidate zonation of the relatively deep-water carbonate–siliceous Jurassic and Cretaceous sequences in the Mediterranean area, for example, of the Majolica Formation in Italy (Baumgartner, 1984; Baumgartner *et al.*, 1995; Jud, 1994). Our study on radiolarians allowed us, on one hand, to get insight into the detailed stratigraphy of Jurassic and Cretaceous deposits of the Ankara region and, on the other, to correlate the Alacaatli section with the sections in the east Mediterranean area. We investigated and sampled in detail the Alacaatli section in 1994, giving particular attention to the siliceous and micritic limestones with visible radiolarian remains, occasionally exposed at the weathered surface. Radiolarians were recovered from the rock samples by dissolving them in diluted hydrochloric acid. When the siliceous radiolarian skeletons were replaced by calcite, we used the concentrated fluoric acid to substitute calcite for fluorite. As a result, many stratigraphic levels yielded representative radiolarian assemblages. In addition, we studied in

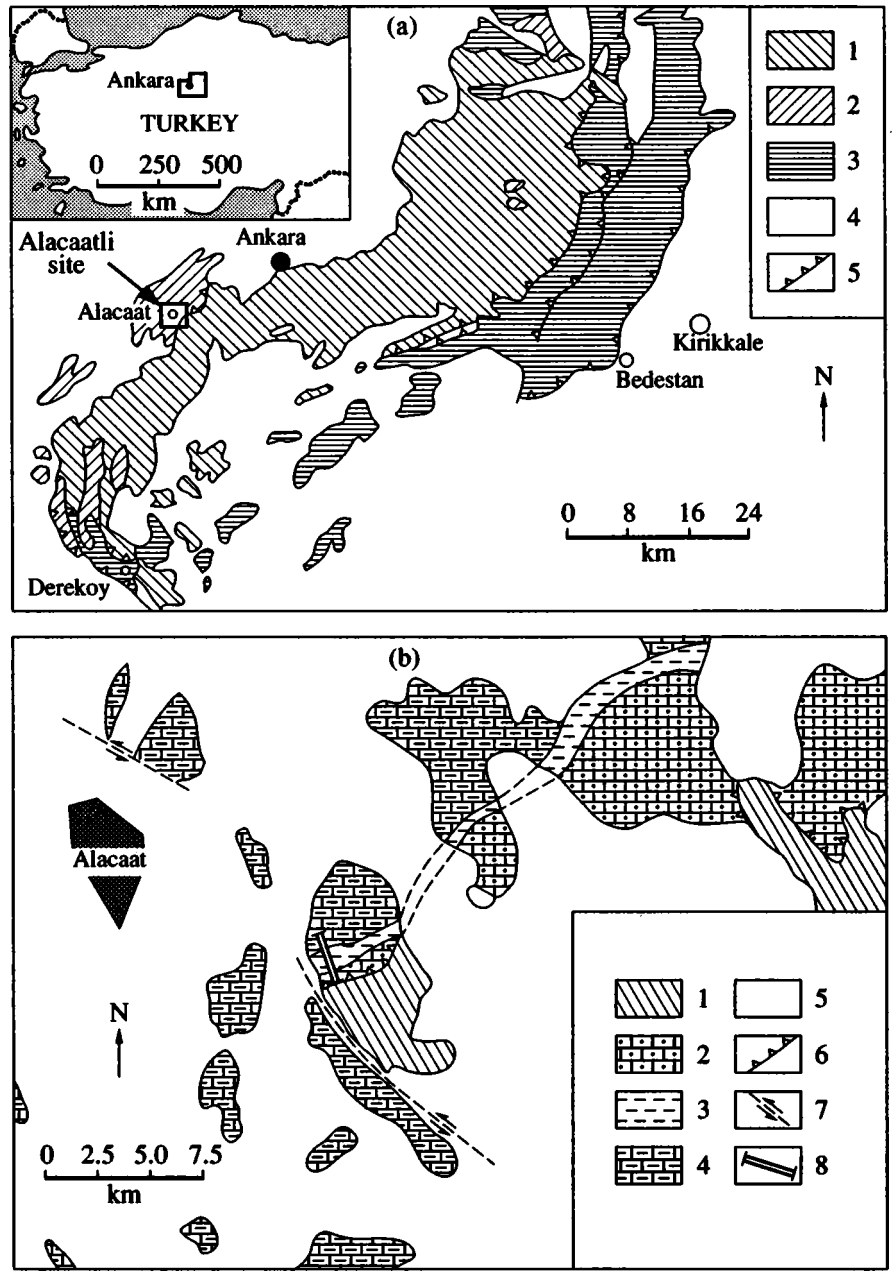


Fig. 1. Position of the Alacaatli section in the structure of the Ankara suture zone. Geological scheme of the Ankara region (A) shows: (1) Paleozoic–Triassic metamorphic shales; (2) Jurassic–Lower Cretaceous deposits; (3) pre-middle Campanian ophiolite melange; (4) middle Campanian and younger deposits; (5) overthrusts.

Geological scheme of the Alacaatli region (B) shows: (1) Paleozoic–Triassic metamorphic shales; (2) middle Oxfordian–lower Tithonian limestones and cherts; (3) upper Kimmeridgian–lower Berriasian marls and limestones; (4) upper Tithonian–upper Valanginian limestones; (5) Upper Cretaceous and younger deposits; (6) overthrusts; (7) strike-sleep faults; (8) position of the studied section.

detail peculiar features of the contact between Jurassic rocks and the underlying Karakaya Formation. The new data provide grounds for making considerable corrections in the stratigraphic zonation and for more accurate interpretation of the regional geologic structure.

To the east of the Alacaatli Village, the following rock units and their relations are exposed at the right-hand wall of the Dericendere gorge (Fig. 2).

Gray massive limestones with Permian foraminifers occur here as a large block in terrigenous melange (the Karakaya Formation).

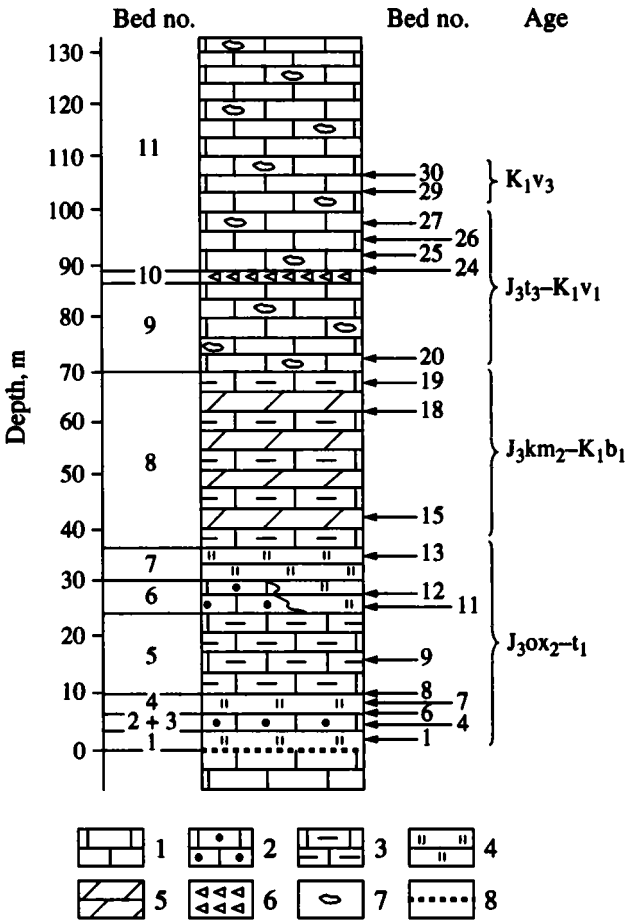


Fig. 2. Lithology of the Alacaatli section: (1) limestone; (2) sandy limestone ("ammonitico rosso"); (3) clayey limestone; (4) chert; (5) marl; (6) sedimentary breccia; (7) cherty concretions; (8) tectonic contact.

The tectonic contact is marked by a thick (5–15 m) zone of black mylonite with numerous small blocks of Permian limestones and Late Jurassic cherts. The fault plane is discordant relative to the attitude of the Jurassic lower beds wedging out toward the east. The tectonic contact is associated with numerous second-order faults.

The site demonstrates the following bed sequence (from the base upward):

Bed 1: calcareous cherts, rose- to yellowish- and greenish gray, platy (Sample 1) with radiolarian remains *Acanthocircus trizonalis trizonalis* (Rust), *Emiluvia sedecimporata* (Rust), *E. salensis* Pessagno, *Mirifusus dianae* (Karrer), *Podocapsa amphitreptera* Foreman, *Ristola altissima* (Rust), *Tetratrabs* sp. cf. *T. bulbosa* (Baumgartner), *Triactoma* sp. cf. *T. blakei* (Pessagno), *Tritrabs* sp. cf. *T. casmaliaensis* (Pessagno), *T. ewingi* (Pessagno), and *T.* sp. cf. *T. exotica* (Pessagno) (Plate I, Fig. 2, and table). The assemblage indicates the middle Oxfordian–early Tithonian age of the bed (Baumgartner *et al.*, 1995). The apparent thickness of the bed is 3 m.

Bed 2: limestine, dark-gray, massive; the bed is 1 m thick.

Bed 3: red to rose-gray grumous limestone, indistinctly laminated, sandy, lacking graded structure and enclosing remains of ammonites and belemnites (Sample 4); the bed yielded foraminifers *Cornuspina intervacare* Azbel, *Paleogaudryina varsoviensis* (Biel. et Poz.), *Eoguttulina simplex* (Terq.), *Ophthalmidium strumosum* (Gumb.), *Ceratolamarckina adiposa* Azbel, *Paulina furssenkoi* Grig., *Epistomina nemunensis* Grig., *E. multialveolata* Grig., *Trocholina transversarii* Paalz., and *Globuligerina oxfordiana* (Grig.). The list implies the middle Oxfordian age of the rock and characterizes the foraminiferal *Ophthamidium strumosum* Zone (determination by K.I. Kuznetsova). The bed thickness is 3 m.

Bed 4: multicolored, banded red–yellow–green, slightly calcareous, platy, enclosing thin (up to 0.5 m) interbeds of gray massive limestone (Sample 6). The bed yielded all radiolarian species found in Bed 1 in association with *Acaeniotyle diaphorogona* Foreman, *Angulobracchia purisimaensis* (Pessagno), *Emiluvia orea* Baumgartner, *Napora bukryi* Pessagno, and *Spongocapsula perampla* (Rust). Starting from the middle of the bed (Sample 7), species *Tetratrabs* sp. cf. *T. bulbosa* Baumgartner disappears. These data provide grounds to attribute Bed 4 to the late Oxfordian–early Tithonian. The bed is 4 m thick.

Bed 5: limestone, clayey, greenish-gray, (Sample 9) platy; the unit marks the first occurrence level of radiolarian species *Acaeniotyle umbilicata* (Rust). The bed is 15 m thick.

Bed 6: limestone, red to rose, grumose, clayey, slightly arenaceous (Sample 12), with numerous remains of ammonites, belemnites and foraminifers : *Globuligerina oxfordiana* (Grig.) of the early–middle Oxfordian (determination by K.I. Kuznetsova). The limestone lens is up to 7 m thick and substituted along the strike for chert (Sample 11) with the same late Oxfordian–early Tithonian radiolarian assemblage as in Bed 5.

Bed 7: calcareous chert, reddish green, striped, platy (Sample 13), with the same radiolarian assemblage as in Bed 5. This is the last occurrence level of radiolarian species *Emiluvia sedecimporata* (Rust), *E. orea* Baumgartner, *Napora bukryi* Pessagno, *Spongocapsula perampla* (Rust), *Triactoma* sp. cf. *T. blakei* (Pessagno), and *Tritrabs* sp. cf. *T. casmaliaensis* (Pessagno). The bed is 7 m thick.

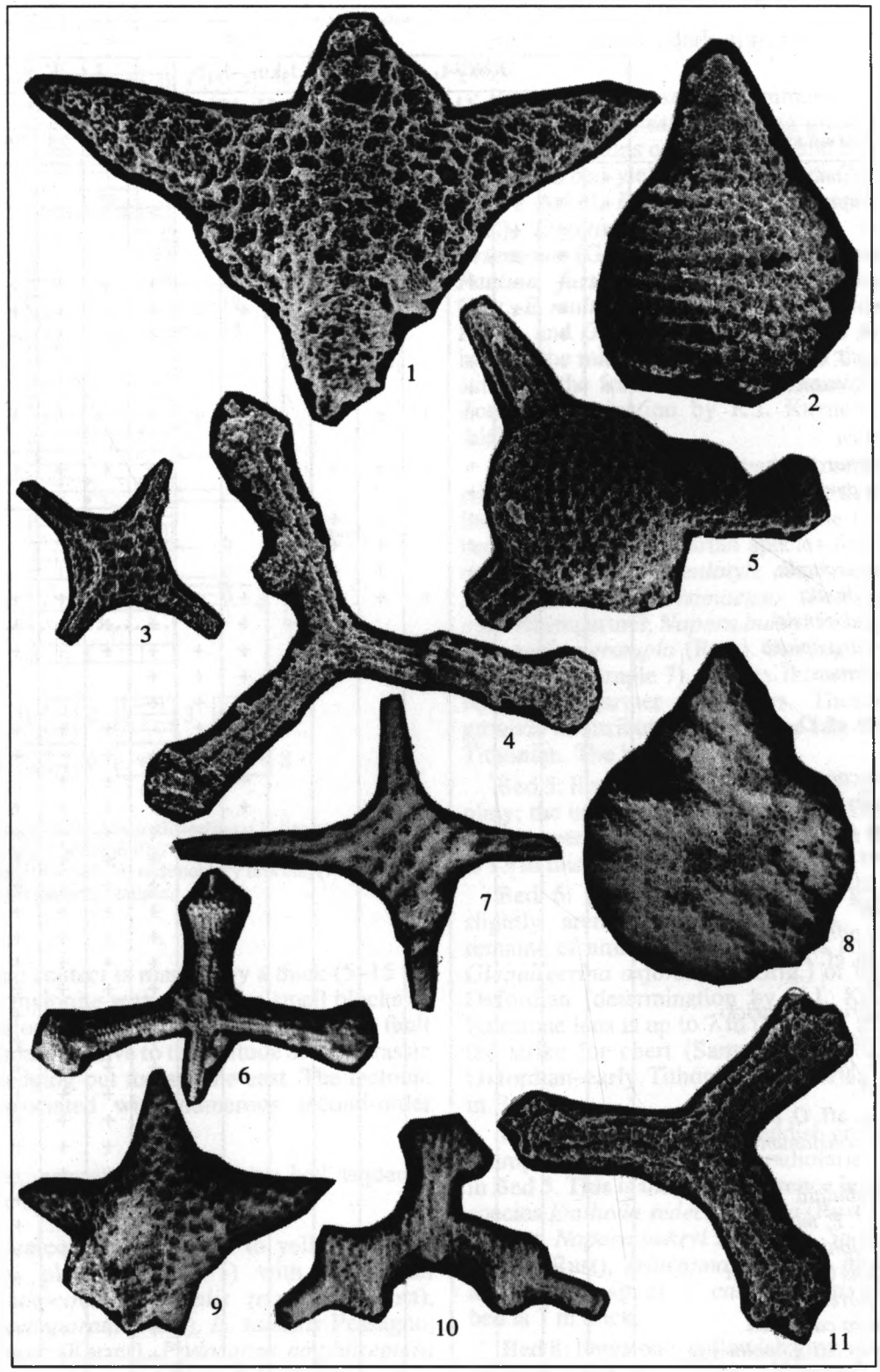
Bed 8: limestone, yellowish gray, clayey, platy, with frequent interbeds of marls, calcareous light-gray claystones and occasional beds of fine-grained calcarenites (Sample 15); radiolarians *Ditrabs sansalvadorensis* (Pessagno) appear at the base of the bed. The upper part of the bed (Sample 18) contains *Archaeodictyomitra apiara* (Rust), *Bistarkum brevilatum* Jud, *B. valdorbi-*

Distribution of radiolarian species in the Alacaatli section

Radiolarian species	Age																
	J_3ox_2–t_1							J_3km_2–K_1b_1			J_3t_3–K_1v_1					K_1v_3	
	9–11*							11–13*			13–16*					17*	
	1	6	7	8	9	11	13	15	18	19	20	24	25	26	27	29	30
Acanthocircus trizonalis trizonalis	+	+	+	+		+											
Emiluvia sedecimporata	+	+	+	+		+	+										
E. salensis	+	+	+	+													
Mirifusus dianae	+	+	+	+	+	+	+		+								
Podocapsa amphitreptera	+	+	+	+		+	+	+	+	+	+	+	+	+	+	+	+
Ristola altissima	+	+	+	+			+	+	+	+	+		+	+			
Tetratrabs sp. cf. *T. bulbosa*	+	+	+							+	+	+	+	+	+		
Triactoma sp. cf. *T. blakei*	+	+	+	+		+	+										
Tritrabs sp. cf. *T. casmaliaensis*	+		+	+		+	+										
T. ewingi	+	+	+	+	+	+	+		+	+	+		+	+	+		+
T. sp. cf. *T. exotica*	+			+													
Acaeniotyle diaphorogona		+	+	+	+	+	+	+	+		+	+	+	+	+	+	+
Angulobracchia purisimaensis		+	+	+													
Emiluvia orea		+	+	+		+	+										
Napora bukryi		+	+	+		+	+										
Spongocapsula perampla		+	+	+		+	+										
Acaeniotyle umbilicata					+	+	+	+	+	+	+	+	+	+		+	+
Ditrabs sansalvadorensis								+	+	+	+	+	+	+	+		
Archaeodictyomitra apiara									+	+	+	+	+	+	+	+	+
Bistarkum brevilatum									+	+	+						
B. valdorbiense									+	+	+						
Obesacapsula sp. cf. *O. morroensis*									+	+		+	+	+			
O. cetia									+	+	+	+		+	+		
O. sp. aff. *O. rusconensis*									+		+	+	+				
Parvicingula boesii									+	+	+	+	+	+	+		
Wrangellium sp. aff. *W. depressum*									+	+	+	+	+	+			
Xitus spicularius									+	+	+	+	+	+			+
Acanthocircus trizonalis dicranacanthos											+	+	+	+	+		+
Alievium helenae											+	+	+	+		+	+
Angulobracchia portmanni											+	+	+	+	+	+	+
Acanthocircus sp. cf. A. hueyi											+	+	+	+			
A. multidentatus											+	+	+	+			
Archaeospongoprunum patricki											+	+	+	+	+		+
Emiluvia chica decussata											+	+	+	+		+	+
Godia sp. aff. *G. concava*											+	+	+	+	+		
Katroma millotti											+	+	+	+			+
Obesacapsula sp. aff. *O. bullata*											+	+	+	+			
Pantanellium berriasianum											+	+	+	+			
P. squinaboli											+	+	+	+			
Paronaella (?) tubulata											+	+	+	+	+		+
Parvivacca sp. cf. *P. magna*											+	+	+	+			
Podobursa polylophia											+	+	+	+	+	+	+
Pseudoaulophacus (?) *pauliani*											+	+	+	+			
P. (?) sp. cf. *P. florealis*											+			+			
Pseudodictyomitra carpatica											+	+	+	+	+		+
Spongocapsula sp. aff. *S. palmerae*											+	+	+	+			
Syringocapsa sp. aff. *S. coronata*											+	+	+				
S. longitubus											+	+	+	+	+		
Wrangellium sp. cf. *W. puga*											+	+	+	+	+	+	+
Xitus sp. cf. *X. alievi*												+	+	+	+		+
Cecrops septomporatus																+	+

* Zones (after Baumgartner *et al.*, 1995); 1–30—sample numbers.

Plate I



Radiolarians of the Upper Jurassic and Lower Cretaceous from the Alacaatli section.

Plate I. (1, 9) *Podocapsa amphitreptera* Foreman: (1) Sample 1, ×235; (9) Specimen 7, ×100. (2) *Mirifusus dianae* (Karrer), Sample 7, ×125. (3, 7) *Emiluvia sedecimporata elegans* Pessagno: (3) Sample 6, ×90; (7) Sample 7, ×100. (4) *Tritrabs ewingi* (Pessagno), Samples 19-6, ×90. (5) *Triacoma* sp. cf. *T. tithonianum* (Rust), Samples 19-7, ×145. (6) *Tetratrabs bulbosa* Baumgartner, Samples 19-7, ×40. (8) *Obesacapsula cetia* (Foreman), Samples 19-7, ×110. (10) *Homoeoparonaella (?)* sp., Samples 19-6, ×100. (11) *Tritrabs casmaliaensis* (Pessagno), Samples 19-11, ×140.

ense Jud, *Obeacapsula* sp. cf. *O. morroensis* Pessagno, *O. cetia* (Forman), *O.* sp. aff. *O. ruscinensis* Baumgartner, *Parvicingula boesii* (Parona), *Wrangellium* sp. aff. *W. depressum* (Baumgartner), and *Xitus spicularius* (Aliev) (Plate II). Here (Sample 18), we also detected the last occurrence of *Emiluva salensis* (Pessagno). These data indicate that the bed may range in age from the late Kimmeridgian till the beginning of the Berriasian (Baumgartner *et al.*, 1995). The bed is 33 m thick.

Bed 9: limestone, light-gray to white, pelitomorphic, a little platy, with numerous cherty concretions and interbeds; the bed (Sample 20) contains the radiolarian assemblage of the previous unit and marks the first occurrence level of radiolarian species *Acanthocircus trizonalis dicranacanthos* (Squinabol), *Alievium helenae* Schaaf, and *Angulobracchia portmanni* Baumgartner. The last occurrence level of *Bistarkum brevilatum* Jud and *B. valdorbiense* Jud is recorded here. This implies the age of the bed to be within the period from the late Tithonian to late Berriasian. The bed thickness is 17 m.

Bed 10: calcareous breccia with angular and semirounded detritus of limestone, cherty concretions, calcarenite, and limy sandstone; the cement is calcic and scanty. The thickness is 2 m.

Bed 11: white limestone, pelitomorphic, massive or platy, occasionally bearing shell detritus and isolated ammonite aptychi; the bed incorporates frequent interbeds of gray calcarenite and cherty concretions. There are also thin (up to 1 m) breccia interbeds and structural marks of subaqueous slumping. At the bed's base (Samples 24, 25, 26), there is the first occurrence level of radiolarians *Acanthocircus* sp. cf. *A. hueyi* (Pessagno), *A. multidentatus* (Squinabol), *Archaeospongoprunum patricki* Jud, *Emiluvia chica decussata* Steiger, *Godia* sp. aff. *G. concava* (Li et Wu), *Katroma millotti* Schaaf, *Obesacapsula* sp. aff. *O. bullata* Steiger, *Pantanellium berriasianum* Baumgartner, *P. squinaboli* (Tan), *Paronaella* (?) *tubulata* Steiger, *Parvivacca* sp. cf. *P. magna* Jud, *Podobursa polylophia* Foreman, *Pseudoaulophacus* (?) *pauliani* Jud, *P.* (?) sp. cf. *P. florealis* Jud, *Pseudodictyomitra carpatica* (Lozyniak), *Spongocapsula* sp. aff. *S. palmerae* Pessagno, *Syringocapsa* sp. aff. *S. coronata* Steiger, *S. longitubus* Jud, *Wrangellium* sp. cf. *W. puga* (Schaaf), and *Xitus* sp.cf. *X. alievi* (Foreman) (Plates III–IV). These species indicate the late Berriasian–early Valanginian age of the bed (Baumgartner *et al.*, 1995). Upsection, 15 m from the base of the bed (Sample 29), *Cecrops septemporatus* (Parona) appears, whereas *Pantanellium berriasianum* Baumgartner and *Obesacapsula cetia* (Foreman) disappear, thus implying the late Valanginian age of this level. The apparent thickness of the bed is 40 m.

Consequently, the Jurassic–Cretaceous deposits in the Alacaatli region are tectonically placed over the Karakaya Formation, and the Lower to Middle Jurassic rocks are entirely missing in this section. This is indirectly confirmed by the fact that the reliably dated Lias and Dogger deposits, which transgressively overlie the Karakaya Formation near the Kosrelik Village (10 km north of Ankara), are completely different in lithology (conglomerates, sandstones, and siltstones with interbeds of marls and "ammonitico rosso" facies). On the basis of radiolarians, the following stratigraphic units can be distinguished in the Alacaatli region: the middle Oxfordian–lower Tithonian (beds 1–7), the upper Kimmeridgian–lower Berriasian (Bed 8), the upper Tithonian–lower Valanginian (Bed 9 to the lower part of Bed 11), and the upper Valanginian (the upper part of Bed 11). This conclusion is tentative and requires further discussion and supplementary information.

The composition of radiolarian assemblage from Bed 1 (Sample I) can be assumed as typical of zones 9 through 11 of the Middle Jurassic–Lower Cretaceous standard radiolarian zonation for the Mediterranian region (middle Oxfordian–Lower Tithonian; Baumgartner *et al.*, 1995). It should be noted here that the zonation units are sufficiently detailed, but nonetheless, their stratigraphic intervals sometimes overlap, as one can see from stratigraphic inferences on the section studied.

The age of the foraminiferal assemblage from limestone Bed 3 (Sample 4) is middle Oxfordian, whereas the radiolarian assemblages from underlying (Sample 1) and overlying (Sample 6) siliceous deposits are the middle Oxfordian–early Tithonian and, respectively, late Oxfordian–early Tithonian in age. This age discordance between radiolarian and foraminiferal assemblages can be accounted for in two ways. On the one hand, the Upper Jurassic radiolarian zonation is now under elaboration and revision, and, consequently, there could be certain ambiguities concerning the zonal boundaries currently accepted. On the other hand, it should be taken into consideration that radiolarians were found in siliceous rocks prevailing in the lower part of the Alacaatli section, whereas the foraminiferal finds are confined to the limestone lenses of the "ammonitico rosso" facies of a shallower origin. The conformable bedding of limestones and siliceous rocks, along with the complete absence of small fragments or clastic trains of these carbonate rocks along the strike of limy lenses, provides no grounds to consider the "ammonitico rosso" limestone bodies as slided blocks. The composition of the foraminifera assemblage lacks both the redeposited or untypical specimens (K.I. Kuznetsova, oral communication). This and also the of textural and bedding peculiarities rule out a possibility to explain the situation by a hiatus in sedimentation and redeposition of sedimentary material by turbidity or contour currents. As is generally known, the "ammonitico rosso" facies are characterized by a rather slow lithification. It is not improbable, however, that these limestone bodies represent a result of sliding of nonlithified carbonate deposits and their subsequent burial within the siliceous sediments. To a certain extent, this is implied by the laminated structure of limestones and by

Plate II

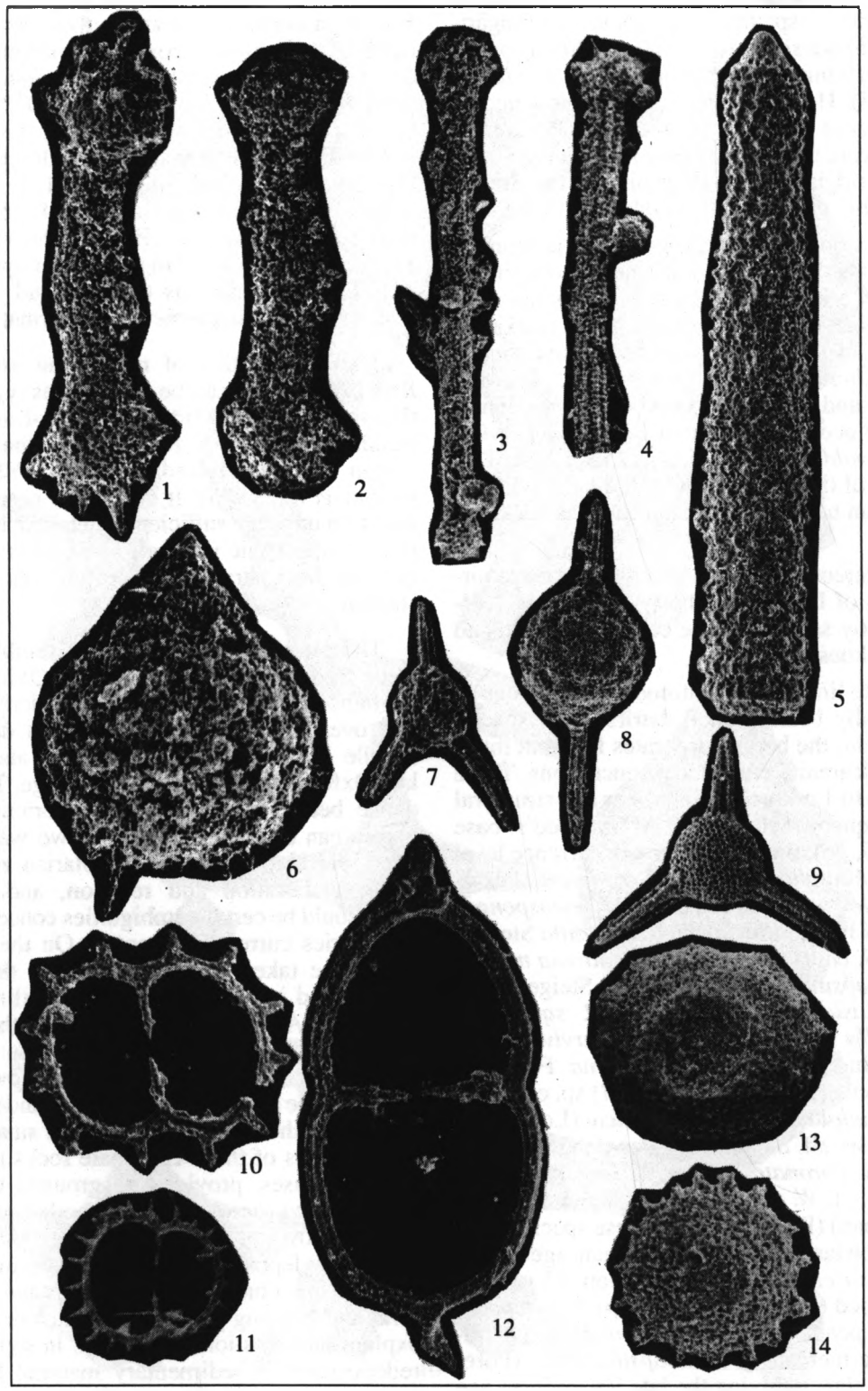


Plate II. (1, 2) *Bistarkum brevilatum* Jud both, Samples 19-18, ×80. (3, 4) *Ditrabs sansalvadorensis* (Pessagno) both; Samples 19-20, ×70. (5) *Ristola altissima* (Rust), Samples 19-25, ×110. (6) *Sethocapsa* sp. cf. *S. trachyostraca* Foreman, Samples 19-18, ×95. (7) *Acaeniotyle diaphorogona* Foreman, Samples 19-25, ×90. (8) *Acaeniotyle unbilicata* (Rust), Samples 19-25, ×90. (9) *Triactoma* sp. ex gr. *T. tithonianum* (Rust), Samples 19-25, ×105. (10) *Acanthocircus* sp. cf. *hueyi* (Pessagno), Samples 19-25, ×130. (11) *Acanthocircus multidentatus* (Squinqbol), Samples 19-26, ×125. (12) *Acanthocircus trizonalis* (Rust), Samples 19-25, ×85. (13) *Godia* sp. aff. *G. concava* (Li et Wu), Samples 19-26, ×140. (14) *Patellula helios* (Squinabol), Samples 19-26, ×140.

Plate III

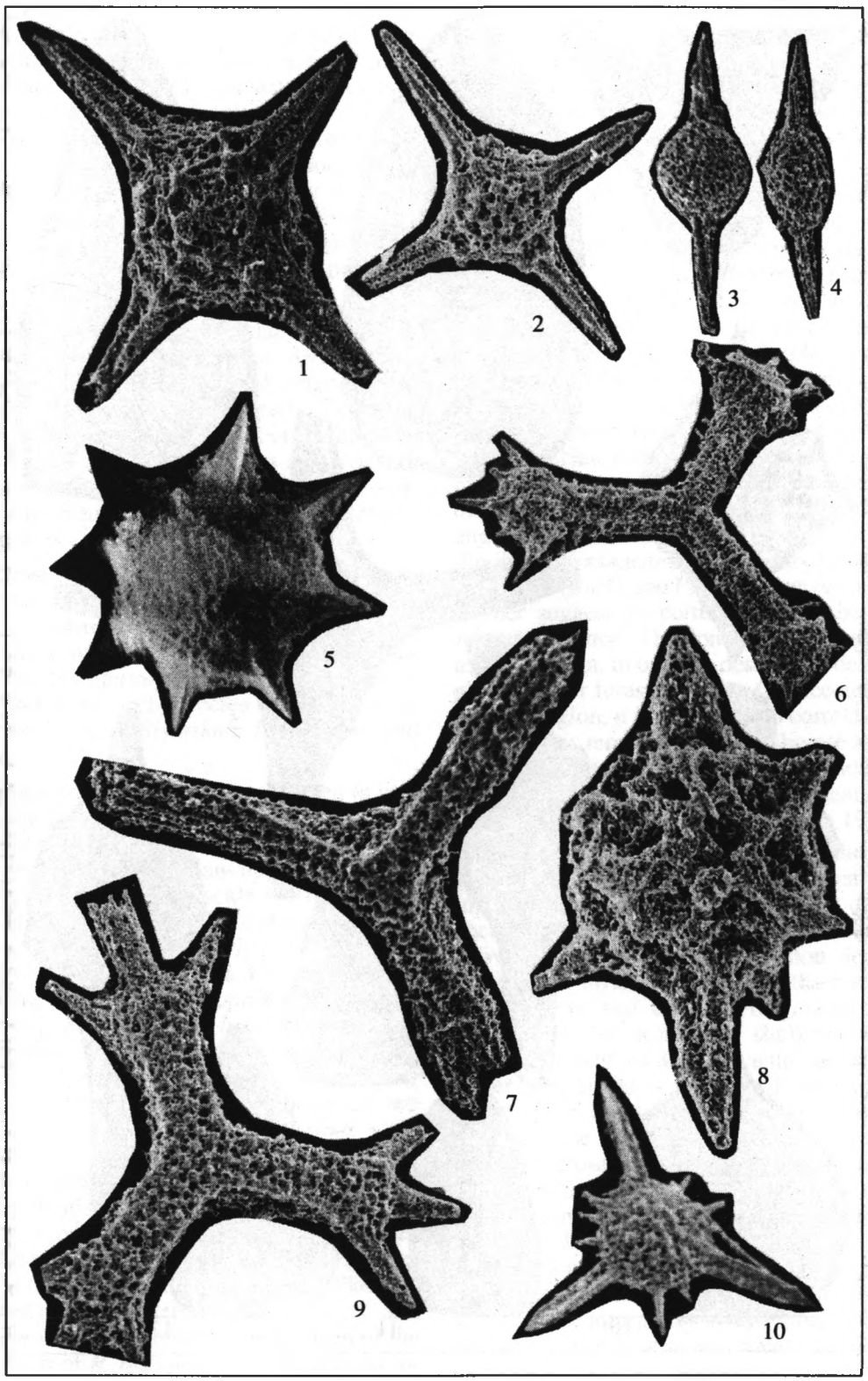


Plate III. (1, 2) *Emiluvia chica decussata* Steiger., Samples 19-25, ×225; and 19-26, ×140. (3) *Archaeospongoprunum patricki* Jud, Samples 19-25, ×120. (4) *Archaeospongoprunum* sp., Samples 19-25, ×120. (5) *Pseudoaulophacus (?) pauliani* Jud, Samples 19-25, ×100. (6) *Tritrabs ewingi* (Pessagno), Samples 19-25, ×90. (7) *Angulobracchia (?) portmanni* Baumgartner, Samples 19-25, ×165. (8) *Pantanelliun berriasianum* Baumgartner, Samples 19-25, ×225. (9) *Halesium (?)* sp. aff. *H. nevianii* (Squinabol), Samples 19-25, ×165. (10) *Alievium helenae* Schaaf, Samples 19-25, ×100.

Plate IV

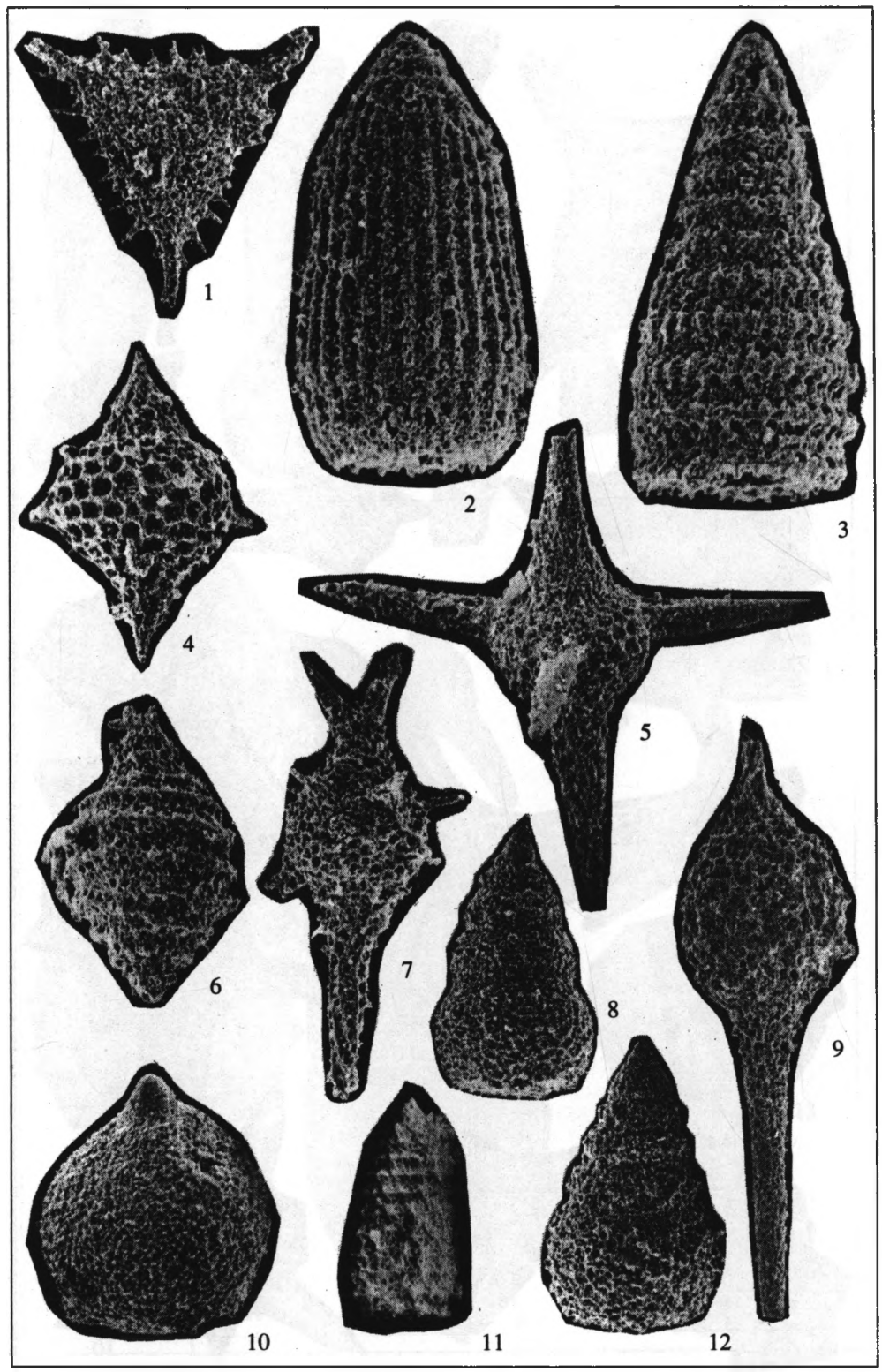


Plate IV. (1) *Paronaella (?)* sp. ex gr. *P. (?) trifoliacea* Ozwoldova, Samples 19-25, ×105. (2) *Archaeodiotyomitra apiara* (Rust), Samples 19-25, ×280. (3) *Wrangellium (?)* sp. cf. *W. (?) puga* (Schaaf), Samples 19-25, ×250. (4) *Syringocapsa coronata* Steiger, Samples 19-25, ×120. (5) *Podocapsa* sp., Samples 19-25, ×100. (6) *Parvicingula boesii* (Parona), Samples 19-25, ×145. (7) *Katroma millotti* Schaaf, Samples 19-25, ×130. (8, 12) *Spongocapsula* sp. aff. *S. palmerae* Pessagno, Samples 19-25, ×100 (9) *Syringocapsa longitubus* Jud, Samples 19-25, ×140. (10) *Obesacapsula* sp. aff. *O. bullata* Steiger, Samples 19-25, ×90. (11) *Xitus* sp. cf. *X. alievi* (Foreman), Samples 19-25, ×100.

the fact that two "ammonitico rosso" horizons that are known in the Alacaatli section are of the same, or very close, early(?)–middle Oxfordian age in spite of their rather different positions in the section. The solution of the problem needs further study on ammonites and belemnites abundant in the "ammonitico rosso" limestones, and on foraminifers and other microfauna in rocks enclosing the limestone blocks. It is noteworthy that the analysis of many thin sections of rocks from the studied site did not reveal the remains of calpionellids, but individual foraminiferal shells were found and could be further studied. Nannoplankton has not been searched yet.

At the level of Bed 4 (Sample 6), many radiolarian new-comers (*Acaeniotyle diaphorogona, Angulobracchia purisimaensis, Emiluvia orea, Spongocapsula perampla*) imply the late Oxfordian–early Tithonian age of beds 4–7, which can be thus correlated with zones 10–11 of the West Mediterranean area (Baumgartner *et al.*, 1995). However, *Globuligerina oxfordiana* from limestone Bed 6 (Sample 12) is characteristic of the lower–middle Oxfordian.

The radiolarian assemblage from Bed 8 considerably differs from assemblages characteristic of the underlying beds and can be correlated with radiolarians from the upper Kimmeridgian–lower Berriasian zones 11–13 of the West Mediterranean region (Baumgartner *et al.*, 1995) owing to such species in common as *Ditrabs sansalvadorensis, Bistarkum brevilatum*, and *B. valdorbiense*.

The assemblage from Bed 9 can be assumed to correspond to the lower–middle Berriasian zone D2 of the Lower Cretaceous scale for West Mediterranean (Jud, 1994), or to the upper Tithonian–upper Berriasian zones 13–14 in the latest zonal scale for this region (Baumgartner *et al.*, 1995) because of the presence of *Acanthocircus trizonalis dicranacanthos, Alievium helenae* and *Angulobracchia portmanni*. The lower part of Bed 11, where *Archaeospongoprunum patricki, Katroma millotti, Pantanellium berriasianum*, and other species appear, is correlative with the upper Berriasian–lower Valanginian zones 14–16 (Baumgartner *et al.*, 1995). In the studied locality, however, we detected radiolarian species *Ristola altissima* s. l., whose distribution, as was indicated in the latest papers (Jud, 1994; Baumgartner *et al.*, 1995), ranges from zone 7 (upper Bathonian) to zone 11 (lower Tithonian). This is the only taxon from the Alacaatli section unfitting the radiolarian zonation presented by Baumgartner *et al.* (1995). These well preserved forms (Plate II; Fig. 5) can hardly be considered as redeposited, and solution of the problem apparently is in the need of the taxonomic revision of *R. altissima* with due regard for its stratigraphic range.

The middle part of Bed 11, judging by the presence of *Cecrop sptemporatus*, can be correlated with zone E2 near the base of the upper Valanginian (Jud, 1994), or with zone 17 of the upper Valanginian (Baumgartner *et al.*, 1995). It is not yet clear how complete is the upper Berriasian–middle Valangiian interval in the Alacaatli section. These beds could have been partly eroded, since the Lower Cretaceous interval of the section includes the intraformational limestone breccias implying the probability of such events, and so the problem can be solved after further study in this direction.

Besides the Alacaatli region, similar Upper Jurassic–Lower Cretaceous deposits were observed to the southwest, near the Ballyukumci Village. Two stratigraphic units are identified here: (1) intercalated beds yellowish gray clayey limestones and marls, whose apparent thickness is up to 40 m; the bed yielded remains of the upper Kimmeridgian–lower Berriasian radiolarians *Bistarkum valdorbiense* Jud and *Sethocapsa cetia* Foreman, thus being correlative with Bed 8 in the Alacaatli section; (2) a unit of light gray and white micritic limestones with interbeds of gray calcarenites; its apparent thickness is about 60 m, and the discovered remains of the upper Tithonian–Lower Valanginian radiolarians *Alievium helenae* Schaaf, *Acanthocircus dicranacanthos* (Squinabol), *Archaeodictyomitra apiara* (Rust), and *Pantanellium berriasianum* Baumgartner suggest its correlation with beds 9–11 of the Alacaatli section. The contact between these units is of tectonic origin. In order to describe in detail the stratigraphy of Upper Jurassic–Lower Cretaceous deposits in the Ankara region, it is necessary to correlate different sections of this territory and to elaborate a zonal scale by applying the unitary association method that was used in the western part of the Mediterranean (Baumgartner, 1984; Jud, 1994; Baumgartner *et al.*, 1995).

It should be noted, concerning the sedimentation conditions in the Alacaatli region, that plankton-bearing carbonate and siliceous deposits dominating here accumulated in an open and sufficiently deep basin above the carbonate compensation depth. Moreover, the section invariably includes the rocks of shallow-water origin and deposits originated as a result of destruction and erosion of shallow-water sediments. The lower part of the Alacaatli section (Oxfordian–lower Tithonian) is composed only of "ammonitico rosso" limestones, which, as mentioned above, may represent the nonlithified sediments that slided down the slope from a shallow-water level. The middle part of the section (upper Kimmeridgian–lower Berriasian) contains calcarenite horizons presumably corresponding to the distal calcareous turbidites. The upper part of the section (upper Tithonian–Valanginian) reveals many beds of calcarenite and carbonate breccia. The latter can be interpreted as proximal calcareous turbidites, the products of active water erosion of the nearby carbonate platform. Sediments of the studied section were apparently accumulated at the base of the carbonate platform slope, in close vicinity to the shelf. They cannot be considered as pelagic deposits of an open ocean. This is evident from the relatively large thickness (about 130 m) of the section that spans the strati-

graphic interval (Oxfordian–Valanginian) only slightly exceeding that of a normal stage. For example, the thickness of Triassic (Olenekian–Rhaetian) siliceous deposits in the southern part of Sikhote-Alin is not much greater than 60 m (Bragin, 1991).

The main conclusions obtained as a result of this analysis are as follows:

(1) The stratigraphic interval of the Alacaatli section spans the Upper Jurassic and basal Lower Cretaceous but not all three series of the Jurassic system, as was supposed by other researchers. Siliceous and cherty limestones with radiolarians are apparently untypical of the Lias and Dogger of the Ankara region.

(2) The radiolarian assemblages from the Alacaatli section are similar to coeval assemblages in the West Mediterranean region, thus providing the possibility to correlate the sections and apply the unitary association method to the Jurassic–Cretaceous zonation in Turkey.

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Albian–Late Cretaceous Phytostratigraphy and Floral Evolution in the North Pacific[1]

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Abstract—Different views on the composition and stratigraphic succession of the Cretaceous floras in northeastern Asia are considered. These floras populated different landscapes of that time and their correlation are difficult to correlate. In the Late Cretaceous time, northeastern Asia and Alaska were partly occupied by sea basins bordered by coastal plains with well developed river systems. In the inner land areas, there were uplands and plains, which escaped flooding. Continental deposits were accumulated there in isolated depressions. The upland of the Okhotsk–Chukchi volcanogenic belt extended along the eastern coast of the Asian continent. All late Albian–Late Cretaceous floras in coastal plains of northeastern Asia and Alaska were dominated by angiosperms. The intracontinental depressions and rises of the volcanogenic upland were occupied by the "Cenophytic" floras with prevailing angiosperms, which are correlatable with the coastal floras, and by the "Mesophytic" ones consisting of dominant ferns and gymnosperms associated with sporadic angiosperms. The coexistence of different floras can be explained by the ability of early angiosperms to occupy the disturbed vegetation-free habitats of the coastal plains, which permanently originated in response to sea transgressions and river meandering. These plant associations gradually penetrated along river valleys into the inner areas populated by coniferous–fern vegetation.

Key words: *North Pacific, Albian, Late Cretaceous, phytostratigraphy, florogenesis, non-marine deposits, landscapes, angiosperms.*

INTRODUCTION

The Cretaceous nonmarine deposits are widespread in the North Pacific province comprising the Arctic areas (those to the north of 65°N paleolatitude) of northeastern Asia and Alaska, which framed the ancient Pacific Ocean on the north. The deposits are usually subdivided and correlated by means of paleobotanic data, whose significance for the geological mapping and stratigraphic investigations in the region is difficult to overestimate.

Intensive studies of the North Pacific Cretaceous floras began in the first half of our century. The most important data were obtained by A.N. Krishtofovich and A. Hollick. Later, the Cretaceous floras and phytostratigraphy of the flora-bearing deposits of the region were discussed in publications of T.N. Baikovskaya, L.Yu. Budantsev, V.F. Belyi, V.A. Vakhrameev, N.D. Vasilevskaya, L.N. Abramova, L.B. Golovneva, A.F. Efimova, V.A. Krasilov, E.L. Lebedev, V.A. Samylina, Ch.J. Smiley, R.A. Spicer, G.G. Filippova, S.V. Shchepetov, and others. I have been involved in investigations of the Cretaceous floras and stratigraphy of the region since the early 1980s.

[1] Report made at the conference in memory of V.A. Vakhrameev in the Geological Institute of Russian Academy of Sciences, Moscow, November 13–14, 1996.

At the beginning of this article, the existing views on the Cretaceous flora development and phytostratigraphic schemes for northeastern Asia are briefly reviewed. The second part of the article presents my approach to the problems, which is based on both the known evidence previously discussed in publications and newly obtained data. I consider as most important for the North Pacific Late Cretaceous florogenesis the relatively rapid occupation of unstable disturbed habitats of the coastal plains by angiosperms and their gradual penetration along the river valleys into the Asian continent. A role of these plants in the Late Cretaceous floral evolution in the region is considered in the third part of the article. Finally, some stratigraphic inferences are suggested.

DIFFERENT VIEWS ON THE CRETACEOUS FLORAS OF NORTHEASTERN ASIA AND THEIR STRATIGRAPHIC SUCCESSION

The first serious attempt to elucidate the succession of the Cretaceous floras in the Pacific coast of Russia was made by Vakhrameev (1966), who analyzed all known Cretaceous floras of the region in terms of their stratigraphic range and composition. Paleobotanical data has been widely used for subdivision and correlation of the flora-bearing deposits of northeastern Russia since the late 1960s, when Samylina (1974) suggested

Table 1. Stages in the Aptian–Campanian development of flora in northeastern Russia, according to different authors

A

Age	Samylina, 1974, 1976, 1988 Stratofloras	Belyi, 1982 Stratofloras	Lebedev, 1987 Horizons
Campanian			Ust'-Emuneret
			Ayanka
Santonian			
Coniacian	Valizhgen		Ketanda
Turonian	Grebenka		
		Grebenka	Dukchanda
Cenomanian	Arkagala		
		Arkagala	Amka
Albian	Arman		Arinda
		Arman	
	Toptan	Toptan	Emanra
	Buor-Kemyus	Buor-Kemyus	Yadrina
Aptian	Silyap		
		Silyap	

B

Filippova and Abramova, 1993 Floral assemblages	Age
Barykov	K_2km
Anadyr'	K_2st
Tyl'pegyrgynai	K_2k
Arkagala	K_2t
Chauna	
Grebenka and Arman	K_2s
Toptan	K_1al
	K_1ap

a phytostratigraphic scheme for the Lower–basal Upper Cretaceous continental deposits. She established eight successive stages of paleoflora development (stratofloras in her terminology) from the Late Jurassic through the Turonian. The scheme adopted at the 2nd Interdepartmental regional conference on northeastern Russia stratigraphy (*Resheniya...*,1978) was widely used in stratigraphic investigations and by geological mapping. Later, some stages were recognized in the regional flora development during the Late Cretaceous (Budantsev, 1970, 1983; Herman, 1985, 1988, 1991; Samylina, 1986, 1988; Kirichkova and Samylina, 1978; Krasilov *et al.*, 1981; Golovneva, 1990; Nesov and Golovneva, 1990) and corresponding Upper Cretaceous phytostratigraphic horizons established in the Anadyr'–Koryak area (Herman, 1993a).

The virtually unmodified scheme of Samylina (1974) was used in some later publications (Belyi, 1982; Filatova and Lebedev, 1982; Lebedev, 1979, 1987, 1992; and others) devoted to the nonmarine Cretaceous stratigraphy of northeastern Russia (Table 1A). According to Samylina, the early angiosperms of northeastern Asia appeared among the Mesophytic vegetation in the early Albian (the Buor-Kemyus stratoflora), increased in abundance and diversity during the Albian, and became predominant for the first time in the Arman stratoflora (the late Albian). However, the succeeding Arkagala stratoflora of the early Cenomanian demonstrates sharply reduced diversity and abundance of angiosperms. In Samylina's opinion, this flora of very peculiar composition (predominantly advanced coniferous genera in association with extremely rare angiosperms) represents a fluctuation of the stratoflora development in response to the extreme cooling during the volcanic activation in the region. From the late Cenomanian (the Grebenka stratoflora), angiosperms became dominants again.

Some researchers (Filippova 1975a, 1979, 1990; Filippova and Abramova, 1993; Krasilov, 1975, 1979; Krasilov *et al.*, 1981) hold an essentially different view on the succession and ages of the Cretaceous flora development in northeastern Asia. They cast doubt on the succession of the "Arman–Arkagala–Grebenka" floras, thus indicating the most vulnerable point in the

Samylina's scheme, namely, the position of the Arkagala stratoflora. Filippova considered the Arman stratoflora as identical in age to the Grebenka flora and suggested the Arkagala stratoflora to be younger than both of them (Table 1B).

The age of the Chauna floral assemblage and its position in the Cretaceous floral succession of northeastern Russia has also been a subject of long debates (Table 1). This assemblage from the volcanic Chauna Group of the northern Okhotsk–Chukchi volcanogenic belt is very peculiar in composition. It shows an abundance of endemic plants and large-leaved cycadophytes, an association of czekanowskialeans with abundant advanced coniferous genera, and a rare occurrence of angiosperms belonging, however, to Late Cretaceous genera. Samylina (1974, 1988) considers this flora to be late Albian in age and contemporaneous with the Arman flora. The same idea is supported by Belyi (1977, 1982, 1994) and Lebedev (1987).

On the contrary, Filippova and Abramova (1993) suggest the early Turonian age of the Chauna flora (Table 1, B). This rises two questions ignored by these researchers. First, why the Grebenka flora with very diverse and abundant angiosperms represented by small- and large-leaved forms was replaced by the Chauna and then by Arkagala floras, whose exclusively small-leaved angiosperms are very rare, although they both were followed by the Tyl'pegyrgynai flora with abundant and mostly large-leaved angiosperms. Second, why the Turonian Chauna and Arkagala floras are concurrent with the sharply different Penzhina flora (Herman, 1991) showing the predominance of large-leaved angiosperms.

Meanwhile, new data concerning the composition and age of the Cretaceous floras of northeastern Asia and casting doubts on some aspects of the schemes of Samylina and Filippova recently became available. The following facts are worth mentioning.

(1) When we studied the Grebenka flora localities and palynological spectra in the right bank of the Anadyr' River, Shchepetov, Belyaya, and myself established an older (late Albian–Cenomanian) age of the flora, than was earlier supposed, and dated fossil plants from the "classic" Grebenka River locality back to the late Albian–early Cenomanian (Shchepetov and Herman, 1990; Shchepetov *et al.*, 1992).

(2) After revision of the Arman flora, Shchepetov and myself supported the earlier supposition of Filippova that it may be similar in age to the Grebenka flora. This is also confirmed by comparison of these two floras with fossil floras of North Alaska (Herman and Spicer, 1997a).

(3) The recent $^{40}Ar/^{39}Ar$ dating of four samples from the volcanics of the upper half of the Chauna Group (the Pykarvaam, Voron'ina, and Koekvun' formations) yielded the age interval from 86.7 to 88.9 Ma (the Coniacian). This suggests that all Chauna deposits and flora can have the Turonian–Coniacian age (Herman *et al.*, 1996; Kelley *et al.*, in press). Correspondingly, the "Arkagala" type floras from deposits overlying the Chauna Group should be younger in age than the Coniancian.

(4) The palynological analysis of samples from the coal- and flora-bearing deposits in the Arkagala depression allowed Markevich (1989, 1990) to infer their late Turonian–Santonian age; earlier, Belyi and Samylina (1987) reported on pollen grains of Senonian age from the Pervomaiskii Formation of the Magadan region. Samylina attributed the taphoflora of this formation to the Arkagala stratoflora.

(5) In the fossil flora from the Ola Formation of the Arman–Ola interfluve, which, according to Samylina (1988), shows typical features of the Arkagala stratoflora, Shchepetov recognized species imprint characteristic of the early Campanian flora from the Barykov Formation of the Ugol'naya Bay, which I described as *Macclintockia beringiana* Herman (Herman, 1985, 1991; Herman and Shchepetov, 1997). The occurrence of this species cannot serve as an unambiguous evidence for the same age of the Barykov and Ola formations, but it supports the Senonian (supposedly Campanian) age of the latter formation and plant remains.

(6) Initially, Shchepetov (1991) suggested the identical Cenomanian–early Turonian age for the Chauna and Arkagala floras, but later (Shchepetov, 1995a) he only indicated the wide age ranges for these floras: the Early–Late Cretaceous for the Chauna flora and the Late Cretaceous (except for the terminal period of the epoch) for the Arkagala flora. Analysis of taxonomic compositions of the Chauna and Arkagala floras allowed Belyi (1997) to insist on their florogenetic affinity. He proposed to exclude the latter from "the general succesion of paleofloras of northeastern Asia and the whole Siberian–Canadian region".

(7) Some researchers pointed to the structural–landscape heterogeneity of northeastern Russia since the middle or terminal Albian onward, when the region comprised three types of landscapes: the volcanogenic upland of the continental margin, intracontinental plains in the west and northwest, and coastal plains in the east and southeast (Krasilov, 1975, 1979; Belyi and Samylina, 1987; Herman, 1989, 1993a, 1994; Lebedev, 1976, 1987, 1992; Pokhialainen, 1994; Belyi, 1997; and others). I proposed (Herman, 1989, 1993a, 1994) to consider these floral–paleogeographic areas as the Okhotsk–Chukchi (OCSR), Verkhoyansk–Chukchi (VCSR), and Anadyr'–Koryak (AKSR) subregions, respectively. It was also proposed to elaborate individual phytostratigraphic schemes for each of the subregions in order to correlate them to each other and to the schemes based on distribution of marine mollusks.

Such an attempt was made by Shchepetov (1995a, 1995b), who suggested the correlation schemes of the nonmarine deposits of three subregions (Fig. 1). For the VCSR, he accepted subdivisions established earlier by Samylina (1974), though considered differently the age

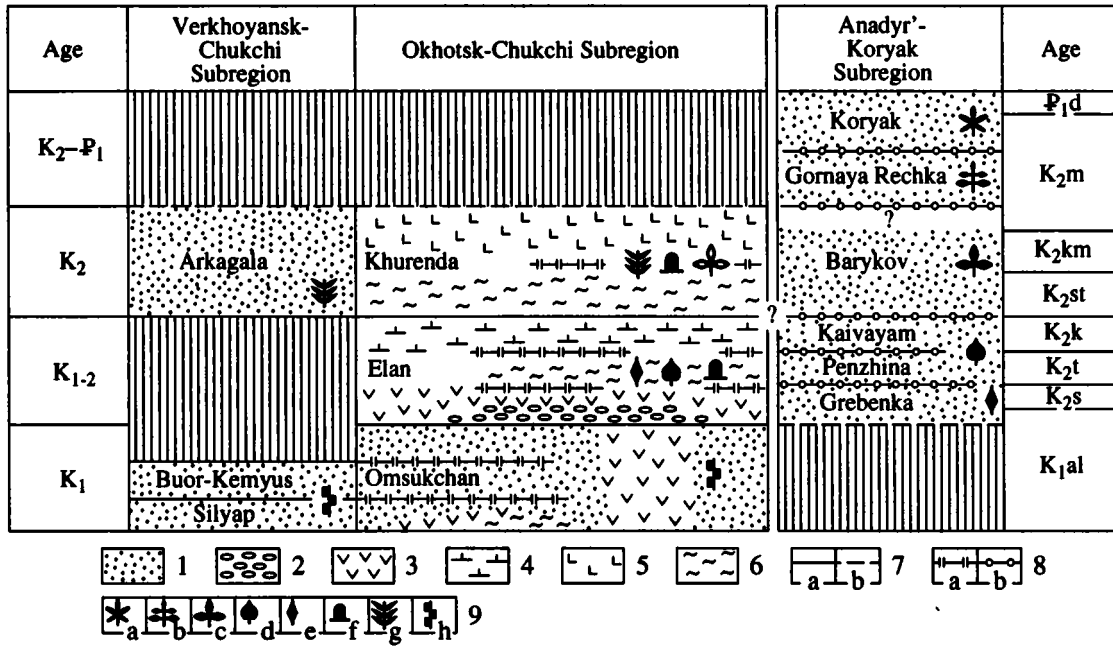


Fig. 1. Correlation scheme for continental Cretaceous in northeastern Russia (Shchepetov, 1995a, 1995b): (1–6), Predominant rocks: (1) sedimentary, (2) volcanogenic–sedimentary, (3) intermediate volcanics, (4) predominantly two-pyroxene intermediate volcanics, (5) predominantly olivine intermediate and basic volcanics, (6) acidic volcanics; (7) established (a) and inferred (b) boundaries of horizons; (8) boundaries of subhorizons (a) and phytohorizons (b); (9) fossil plant assemblages corresponding to the following stages in floral evolution: (a) Koryak (Rarytkin), (b) Gornaya Rechka, (c) Barykov, (d) Penzhina, (e) Grebenka and Arman, (f) Chauna, (g) Arkagala, (h) Toptan, Buor-Kemyus, and Silyap (unfilled symbols denote their probable analogues).

of the Arkagala flora (the Late Cretaceous without the terminal interval). The scheme for the AKSR was that elaborated by myself (Herman, 1993a), but it was partly modified because Shchepetov, as well as Samylina (1986), excluded the Penzhina and Kaivayam stages. For the OCSR, he indicated the coexistence of predominantly coniferous floras of the Chauna and Arkagala types with the angiosperm-dominated plant assemblages analogous to the Grebenka, Penzhina and Barykov floras of the AKSR.

Shchepetov explained the coexistence of these fossil floras from the paleoecological viewpoint: the Chauna type floras were considered to be pioneering communities buried in the tuffaceous Chauna deposits during an epoch of frequent volcanic eruptions, while the Arkagala type floras were interpreted as indicators of "the crisis development of flora in an area of ecological catastrophe" within the volcanogenic belt. The essence of "the crisis-type development" is a permanent reduction of final mature stages of the coenotic succession as a result of volcanic eruptions. In Shchepetov's opinion, the occurrence of the Arkagala flora in the intracontinental Arkagala depression of the VCSR can be explained by the fact that the depression "...was invaded by patients and violents of the crisis succession from the neighboring territory" (Shchepetov, 1995b, p. 22). Shchepetov suggests that the floras analogous to the Barykov flora of the AKSR and rich in angiosperms appeared in the OCSR "...as a result of the intervention or penetration from very competitive species from the coastal plains into the pioneering and succession communities of volcanic plateaus during the volcanic pause" (Shchepetov, 1995b, p. 22).

Using these disputable paleoecological considerations, which should be well substantiated, Shchepetov changed the regional phytostratigraphic scheme suggested by his predecessors for the less detailed one. This actually means a disregard of paleobotanical records used to subdivide and correlate the Cretaceous deposits in northeastern Asia, because the resolution level of the suggested phytostratigraphic scheme (Fig. 1) can hardly satisfy mapping geologists and stratigraphers. There are some questions concerning Shchepetov's interpretation of the Cretaceous flora evolution and the nonmarine Cretaceous stratigraphy of Russia. These unanswered questions, which make his view disadvantageous, are as follows:

What are the grounds for recognizing the abovementioned subregions in the Jurassic–Early Cretaceous (pre-Middle Albian) time, when the respective structural–landscape differentiation has not been developed yet?

Is it possible that all five formations of the Chauna Group (total thickness of 4000 m) contain only the pioneering plant associations, although the initial succession phases usually span a few years to decades, i.e., are instantaneous in the geological sense?

Why were the pioneering (in Shchepetov's opinion) associations dominated by ferns and gymnosperms, i.e., by the plants which are less capable than angiosperms (extremely rare in the Chauna flora) to constitute vegetation of the initial phases of the phytocoenotic succession?

How could "the ecological unstability" in the area of "the active land volcanism" favor "...the conservation of some relict plant forms" (Shchepetov, 1995b, p. 23)?

What are the principles to classify fossil plants into explerents, violents, and patients according to their life strategies that is used by Shchepetov to explain the florogenetic processes in the region?

How could "the permanent reduction of mature phases of the coenotic succession," which, in Shchepetov's opinion, had a considerable impact on evolution, occur over the enormous territory of the OCSR within the geological time frame?

Why did the Arkagala type flora, which was formed in the OCSR "...as a result of long development of plant communities in the unstable (crisis) environment of land volcanism" (Shchepetov, 1995b, p. 23) become preadapted to extremely stable environments of the isolated intracontinental coal-bearing Arkagala depression and spread over it, but could not penetrate into the AKSR and more distant areas of northern Asia and Alaska?

My views on the character of the Cretaceous florogenetic processes in northeastern Asia essentially differ from those of Shchepetov. My interpretation of the Late Cretaceous development of the North Pacific flora (Table 2) is based on data on the North Alaskan fossil floras, which are appropriate for better understanding of the presented views and, as said above, for more precise dating of some North Asian floras.

FLOROGENESIS IN THE NORTH PACIFIC REGIONS DURING THE ALBIAN–LATE CRETACEOUS

In the early and middle Albian time, the vast area from the Lena River basin to Alaska and Western Canada and from the western Okhotsk region to the Arctic Siberian islands was occupied by relatively homogenous but extremely abundant and diverse (150 species) flora of the Buor-Kemyus type (Prinada, 1938; Samylina, 1960, 1964, 1967, 1974, 1976; Lebedev, 1974; Filippova, 1975b; Smiley, 1969a, 1969b; Herman and Spicer, 19997b; and others) (Fig. 2). The early–middle Albian floras of this type were found in the following regions: the Zyryanka locality at the left bank of the Kolyma River (Zyryanka Coalfield); the right bank of the Kolyma River (Omsukchan coal basin); the western Okhotsk region (the Torom depression); the Ula depression; the Bol'shoi Anyui (Ainakhkurgen depression), Eropol (Umkuveem depression), and Penzhina (Verkhnyaya Penzhina and Chalbugchan depressions) river basins; the Kotelnyi Island; the Krest

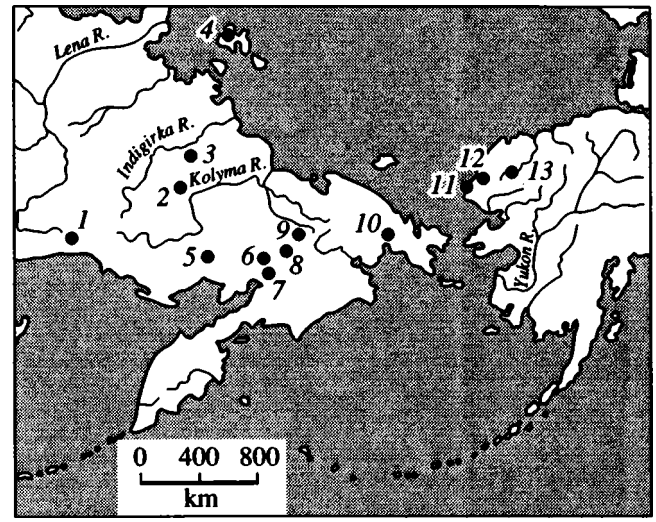


Fig. 2. Main localities of the Buor-Kemyus type flora in the North Pacific region: (*1*) Ula depression; (*2*) Syryanka River; (*3*) Indigirka River basin; (*4*) Kotel'nyi Island; (*5*) interfluve between Balychan and Sugoi rivers; (*6*) Upper Penzhina depression; (*7*) Chalbugchan depression; (*8*) Eropol River basin (Umkuveem depression); (*9*) Bol'shoi Anyui River basin (Ainakhkurgen depression); (*10*) Krest Bay; (*11*) Cape Lisbern; (*12*) Kukpowruk River; (*13*) Corwin River basin.

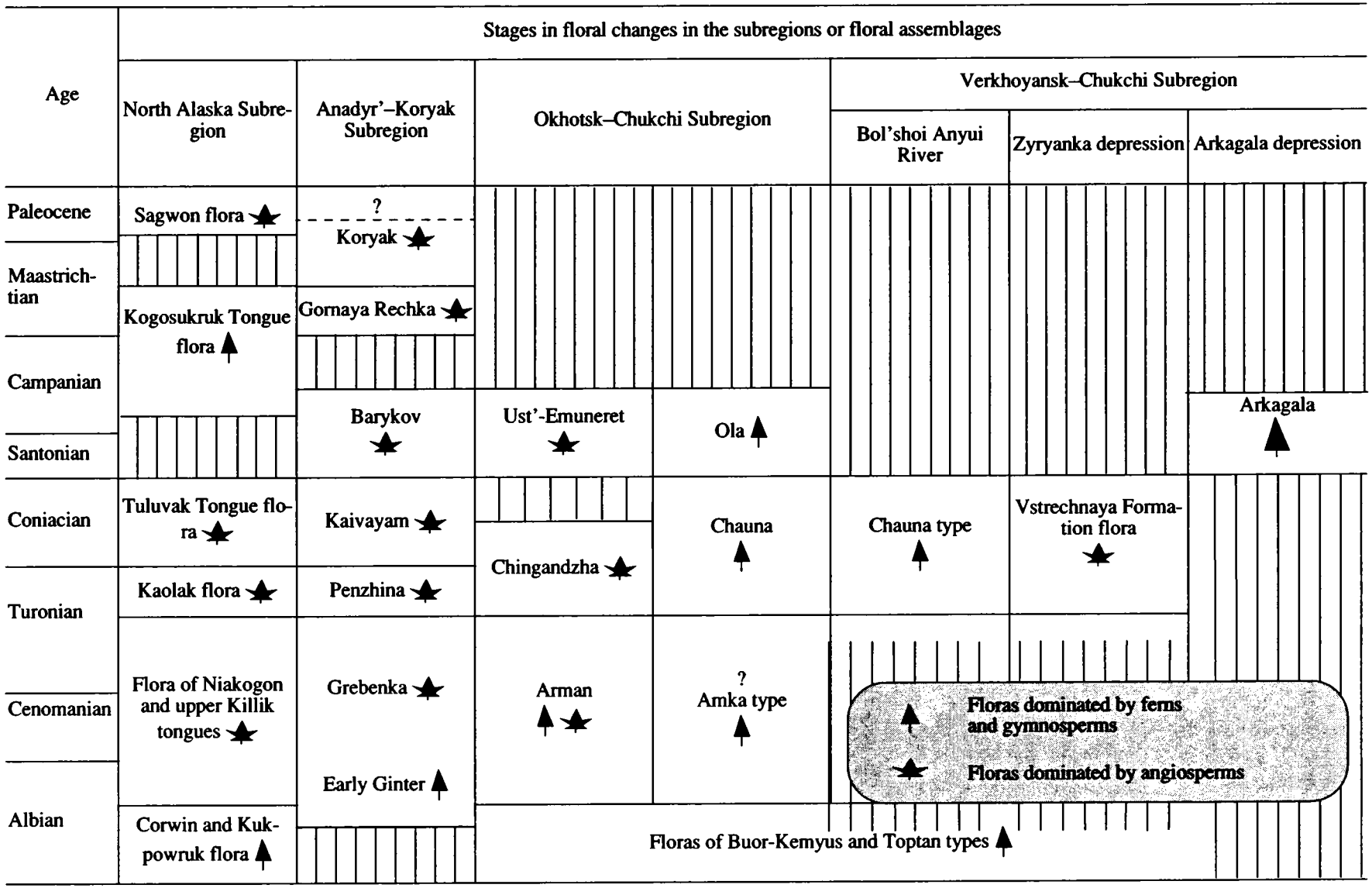
Bay coast, the Cape Lisbern area, and Kukpowruk River basin in the west of North Alaska; the Colville River basin in the east of North Alaska; and Western Canada.

In the North Pacific regions, early angiosperms appeared in the Buor-Kemyus flora. At that time, they were rare and irregularly spread throughout the territory; many of the aforementioned localities lack these plants. The Toptan flora of the Omsukchan coalfield, which represents, in Samylina's opinion (1976), the younger stage of flora evolution in northeastern Asia, may be referred to the Buor-Kemyus type flora. However, because of the homogeneity of the Buor-Kemyus flora, no floral–paleogeographical units of the early–middle Albian time can be recognized in the North Pacific regions and adjacent Siberian–Canadian areas.

In the middle–late Albian time, the North Pacific regions underwent significant paleogeographic changes and radical structural–landscape rearrangement (Belyi and Samylina, 1987; Herman, 1993a; Belyi, 1997): a volcanogenic belt began to form along the eastern margin of Asia, while the Brooks Range continued to elevate in Alaska. This resulted in the disintegration of the North Pacific territory into some floral–paleogeographic subregions differing both in landscape features and composition of floras (Fig. 3). The following continental subregions can be distinguished for the late Albian–Late Cretaceous period (Herman, 1993a).

The Verkhoyansk–Chukchi subregion (VCSR) represented an intracontinental low or hummocky plain

Table 2. The Albian–Paleocene florogenesis in northeastern Asia and North Alaska



escaping submergence, where isolated depressions accumulated nonmarine terrigenous and coaly sediments during the Late Cretaceous.

The Okhotsk–Chukchi subregion (OCSR) embraced the newly formed volcanogenic upland with variously dissected relief. The character of volcanics suggests that "they erupted from volcanoes of fissure and central types, which were 2000–3000 m high and greater" (Vakhrameev, 1989, p. 38). The volcanogenic and terrigenous deposits accumulated in intermontane depressions and volcanic plateaus.

The Anadyr–Koryak subregion (AKSR) was a part of the Kamchatka–Koryak sea basin neighboring the OCSR upland, which emerged at the end of the Albian. The AKSR represented a series of coastal plains with well developed river systems, oxbow-lakes, lakes, and swamps. At some points in the Cretaceous history of the subregion, these plains were flooded by sea and turned into shallow-water basins with abundant islands. At these moments, the continental terrigenous and coal sedimentation was replaced by that of the shallow-water marine type. In the southeast, the AKSR was bounded by the vast Kamchatka–Koryak sea basin; to the north and northwest of the basin there was land, the provenance of terrigenous and volcanogenic material.

The North Alaska subregion (NASR) was similar to the AKSR in the landscape character and in the type of sedimentation. However, there are nonmarine deposits older than the late Albian (early–middle Albian) and bearing the Buor-Kemyus type flora (Herman and Spicer, 1997b; Smiley, 1969a, 1969b). In other words, in the NASR, the late Albian–Late Cretaceous continental sedimentation and flora evolution were inherited from the preceding period, when the botanical–paleogeographical subregions have not been individualized yet. The NASR was bounded by the Arctic sea basin in the north and by the young uplifts of the Brooks Range in the south and southeast. The terrigenous material was supplied from the uplifts, while the tuffaceous sediments of the NASR probably accumulated owing to the Cretaceous volcanic activity in the OCSR (Herman *et al.*, 1996; Kelley *et al.*, in press).

The Yukon–Koyukuk subregion (YKSR) was similar to the AKSR and NASR in the landscape and sedimentation type. However, the nonmarine sedimentation, which succeeded the marine sedimentation of the first half of the Cretaceous, occurred during a very short interval, supposedly, from the Cenomanian (may be since the late Albian) through the Turonian or, locally, early Senonian.

The Chignik subregion (CSR) represented an inextensive coastal plain, which was regularly flooded by sea and existed from the Campanian (supposedly, the late Campanian) till the end of the Paleocene. The Campanian time was marked by the deposition of the Chignik Formation composed of cyclic coastal–marine and coastal–continental terrigenous and coal-bearing beds.

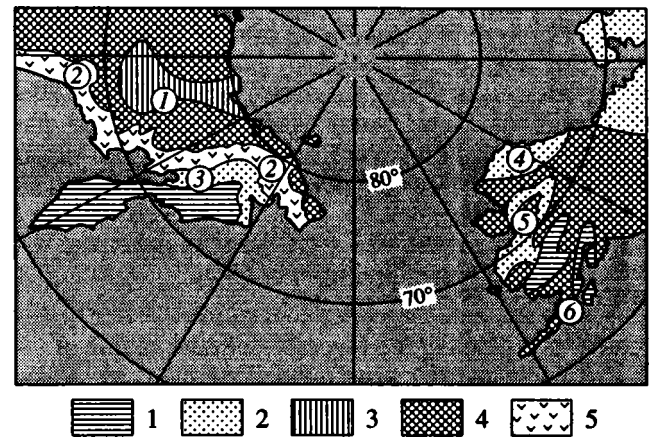


Fig. 3. The Late Cretaceous paleogeography of the North Pacific (after Belyi, 1994, with modifications) and botanical–paleogeographical subregions; (Late Cretaceous position of continents after Smith *et al.*, 1981): (1) areas of stable marine regime; (2) coastal plains repeatedly flooded by sea; (3) plains out of flooding; (4) uplands; (5) volcanogenic upland of the Okhotsk–Chukchi volcanogenic belt; encircled numbers of subregions: (*1*) Verkhoyansk–Chukchi, (*2*) Okhotsk–Chukchi, (*3*) Anadyr'–Koryak, (*4*) North Alaska, (*5*) Yukon–Koyukuk, (*6*) Chignik

Naturally, there were many differences between vegetations of the listed landscapes (Table 2). All the floras of the Late Cretaceous (beginning from the terminal Albian), which existed in the coastal plains (the AKSR, NASR, YKSR, CSR) of northeastern Asia and Alaska, were dominated by angiosperms (Herman, 1993a; Spicer and Parrish, 1990; Hollick, 1930). These floras show distinct evolutionary succession (Herman, 1991, 1993a; Herman and Spicer, 1997a). In the intracontinental VCSR depressions and in the OCSR volcanogenic uplands, slopes, and intermontane depressions, the Late Cretaceous vegetation included both the "Cenophytic" floras dominated by advanced angiosperms, which are well correlated with those of the coastal plains of the AKSR and NASR, and the "Mesophytic" floras predominantly consisting of ferns and gymnosperms in association with sporadic angiosperms; the examples are the late Albian–Cenomanian taphofloras of the "Amka" type (Lebedev, 1987), the Turonian–Coniacian Chauna flora, and the Senonian Ola and Arkagala floras. Noteworthy is that, in the Turonian–Senonian "Mesophytic" floras, the relatively old ferns and gymnosperms, which are relict for the Late Cretaceous, are associated with "advanced," although extremely, rare conifers and angiosperms from genera widespread since the Late Cretaceous only.

The coexistence of the Late Cretaceous floras of different types in northeastern Asia can be explained by the ability of the Cretaceous angiosperms to occupy the disturbed habitats of the coastal plains owing to their reduced reproductive cycle, good protection of ovules, and generalized way of pollination and seed dispersal

(Retallack and Dilcher, 1981a, 1986). In the Late Cretaceous, coastal plains with the developed river system were characteristic of the AKSR and NASR (Koporulin and Voznesenskii, 1969; Sable and Stricker, 1987). The vegetation of these plains was repeatedly and frequently demolished by sea transgressions and locally by meandering river channels and lateral erosion. This permanently generated the disturbed habitats and new vegetation-free substrates after the sea retreat and formation of sandy–pebble meander bars.

Evidently, the angiosperm-dominated associations rapidly spread over the favorable coastal plains of the AKSR and gradually penetrated along the river valleys into the inner areas of the Asian continent populated mainly by the conifer–fern vegetation (Table 2). To some extent, the dispersal of these associations, which preferred disturbed environments and unpopulated substrates, was promoted by the OCSR volcanic activity and climatic fluctuations, which were drastic enough to destroy the vegetation in the high latitude areas and uplands. Most likely, the gradual distribution of the associations dominated by angiosperms over the intracontinental areas of Asia accounts for the coexistence of different-type floras (Table 2) and for the occurrence of ferns and gymnosperms, which are relict for the Late Cretaceous, in the VCSR and OCSR (Vakhrameev, 1988; Samylina, 1988; and others).

The peculiar composition and genesis of the Chauna flora have not been sufficiently elucidated yet and need to be further studied. One cannot exclude that this flora includes some components of the pioneering associations (e.g., *Equisetites* and, probably, some ferns). Another explanation suggests that the specific features of newly deposited ash substrate, such as chemical composition; content of microelements and volatile components, mobility, aeration and drainage properties; and others, could affect the composition of the Chauna flora. In other words, the latter may be supposed to include not only the pioneering plant associations (Shchepetov, 1995b), but also the specialized plant communities adapted to the volcanic ash soils. A well grounded explanation was suggested by Samylina (1974, p. 34), who stated that "...this assemblage represents the mountain vegetation in area of the active volcanism... We are most likely dealing with plant associations, which were not fossilized in usual environments but were rapidly buried under the products of volcanic eruptions."

The Chauna flora may be originated from the older "Amka" floras or Buor-Kemyus flora (Table 2) of the OCSR. The Ola flora (or according to other researchers, the Arkagala type flora of the Okhotsk–Chukchi volcanogenic belt) may have florogenetic connections with the "Amka" and Chauna floras of the subregion. Shchepetov (1995a, 1995b) considers the flora of the Arkagala–Ola type as representing a new stage of the Late Cretaceous flora evolution under unstable conditions of land volcanism in the OCSR. On the contrary, I suppose that the OCSR Ola flora and the VCSR Arkagala flora are the last two "Mesophytic" floras, which had no descendants and "survived" until the Senonian time in the intracontinental areas invaded later than other areas by more advanced plant associations dominated by angiosperms. The taxonomic differences between the Arkagala and Ola floras stem from the fact that the former represents the vegetation of the vast swampy intracontinental depression, while the latter one represents that of mountain slopes, small intermontane depressions, and volcanic plateaus (Belyi and Samylina, 1987). It should be noted that these landscapes themselves were unfavorable for dispersal of the Cretaceous angiosperms: their remains occur sporadically in the pre-Paleogene coal facies everywhere in the region and coniferous vegetation still dominates in the mountain areas.

The post-early Campanian floras are known only from the AKSR and NASR (Table 2). The NASR Campanian–Maastrichtian flora sharply differs from the coeval AKSR floras being of very low taxonomic diversity (eight forms including only two forms of angiosperms: Spicer and Parrish, 1990) and more cryophilic in character. The Campanian–Maastrichtian vegetation of the NASP can be considered most analogous to the recent coniferous taiga of the northern regions. The specific features of this flora may be attributed to the location of North Alaska further to the north (up to 85°N according to some reconstructions) and to the general paleogeographic situation in the northern hemisphere (Herman and Spicer, 1997a). The very impoverished taxonomic composition of the flora under consideration may reflect a significant temperature gradient in the near-polar region during the terminal Cretaceous.

In the taxonomic and physiognomic aspects, the late Maastrichtian Koryak flora of the AKSR (Herman, 1993b; Herman and Spicer, 1995) has much in common with the Paleocene Sagwon flora of the NASR (Spicer *et al.*, 1994). Probably, many taxa of the former flora migrated to North Alaska from northeastern Asia during the Paleocene warming event. The similarity of these floras occurring in both sides of the Cretaceous–Paleogene boundary (Table 2) evidences against the hypothesis of global ecological crisis of this time (Golovneva, 1990, 1994; Spicer *et al.*, 1994). In any case, this crisis had no impact on the evolution of the North Pacific floras. It may be the late Maastrichtian flora of the AKSR that gave rise to the Early Tertiary deciduous floras of the Arctic and adjacent regions (Herman and Spicer, 1995, 1997c).

ROLE OF ANGIOSPERMS IN THE EVOLUTION OF THE ALBIAN–LATE CRETACEOUS FLORAS OF THE NORTH PACIFIC

In northeastern Russia, remains of the oldest (early–middle Albian) angiosperms were mostly found in the Buor-Kemyus Formation of the Zyryanka coalfield (the Kolyma River basin) and in the Tyl' Formation of the

Torom depression (the western Okhotsk region). In both areas, fossil plants occur sporadically and very irregularly throughout the section and are mostly confined to the coarse-grained sandstones of the upper part of the lower Buor-Kemyus Subformation and to the silty sandstones in the upper beds of the lower (conglomeratic) part of the Tyl' Formation (Samylina, 1960, 1967, 1974; Popov, 1962; Lebedev, 1976, 1987), i.e., they are buried in the coarse-grained carbon-free deposits of evidently alluvial–deltaic origin.

Samylina (1967) suggests that the worse preservation of angiosperms of the Buor-Kemyus flora, as compared with ferns, can serve as evidence for the transportation of angiosperm leaves by river currents from the uplands, where, in her opinion, the early representatives of this plant group had their habitat. This point of view, supported later by Lebedev (1976, 1987), is hard to agree with, because the transportation of leaves for a considerable distance by high-energy water currents rich in coarse sandy material would inevitably result in their complete destruction. A more probable interpretation is that the leaves of angiosperms occupying areas near the river channels were rapidly buried in alluvial sands, without being transported for a long distance.

In North Alaska, remains of the early angiosperms are known from the Nanushuk Group (Albian–Cenomanian), which represents the classic post-orogenic mollassa-type deposits characteristic of the deltaic environment. The group is composed of alternating marine (foredeltaic shallow-water) and continental (lagoonal, coastal, and estuarine, as well as deltaic and alluvial) sediments (Sable and Stricker, 1987). The structural differences between the Nanushuk Group sections in the western (the Utukok-Corwin area) and central (the Umiat-Chandler area) parts of North Alaska suggest the coexistence of two large deltaic systems in the Albian–Cenomanian time (Fig. 4); the western system named the Corwin delta and the eastern one known as the Umiat delta (Sable and Stricker, 1987; Affolter and Stricker, 1987). Later (in the early Late Cretaceous?), the Umiat delta was absorbed by the larger and more rapidly developing Corwin delta (Affolter and Stricker, 1987).

In eastern North Alaska, remains of the early angiosperms are encountered in the flora from the upper part of the lower half of the Killik Tongue (Smiley, 1969a; Spicer, 1987; Spicer and Parrish, 1990). This part of the section may correspond to the middle and upper parts of the Corwin Formation in western North Alaska, where the early angiosperms were also found (Smiley, 1969b). Judging from the relationship between the mentioned deposits and marine beds of the Nanushuk Group, the former can be assigned to the middle–basal upper Albian. In 1996, Spicer and myself studied the locality of the Albian angiosperms in the Corwin Formation of the Kukpowruk River basin, western North Alaska (Herman and Spicer, 1997b). There are exposed fluvial sandstones alternating with

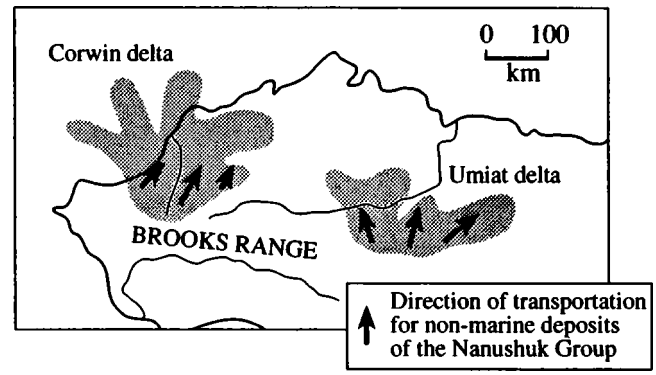


Fig. 4. Outlines of the Corwin and Umiat river deltas of North Alaska in the Albian–Cenomanian time (after Affolter and Stricker, 1987).

siltstones and carbonaceous siltstones of the flood plain. The deposits rest immediately on thick (25–30 m) cross-bedded sandstones of fluvial origin. Therefore, we can state with confidence that in this region, as well as in northeastern Russia, the remains of the early angiosperms occur in the deposits accumulated predominantly under alluvial–deltaic environments of sedimentation.

Thus, the data presented allows me to conclude that the early angiosperms, which appeared in the North Pacific in the early–middle Albian, preferred unstable environments of river flood plains and large river deltas, where they probably occupied newly formed sand bars and channel banks. In the North Pacific, as stated above, the crucial role in the late Albian–Late Cretaceous florogenesis belonged to the relatively rapid dispersal of the angiosperm-dominated associations over disturbed habitats of the coastal plains with well developed river systems and to their penetration along the river valleys into the inner areas of the Asian continent. The paleobotanical data on the Albian–Late Cretaceous North Pacific floras does not confirm the hypothesis that the Early Cretaceous angiosperms inhabited uplands and appeared in geological records in responce to a decrease in the hypsometric level of their habitats (Vakhrameev, 1947, 1952; Arnold, 1947; Samylina, 1974; Lebedev, 1987; and others). On the contrary, my data agrees well with the conclusions of Retallack and Dilcher (1986) that the Cretaceous angiosperms invaded the sea coasts of North America.

The Early Cretaceous angiosperms (as well as, probably, their ancestors and many descendants) could colonize the disturbed habitats because of their ability to be pollinated by wind and insects and to disperse the abundant small seeds with water and wind (Dilcher, 1979; Retallack and Dilcher, 1981a, 1981b; and others). So, being relatively independent from local animals and plant communities, the angiosperms occupied earlier than other plants (due to their reduced reproductive cycle; Stebbins, 1974) the disturbed and new vege-

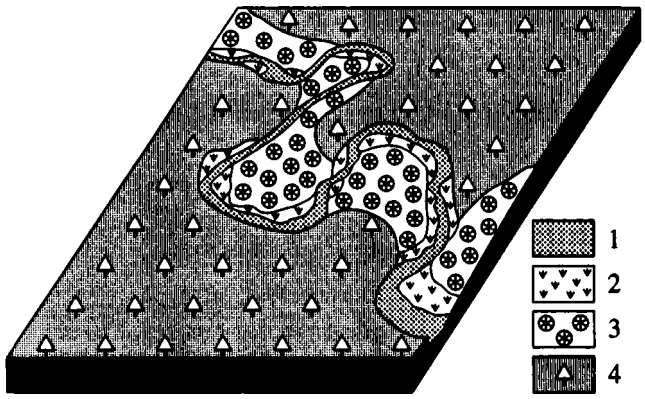


Fig. 5. Vegetation in the valley of a small river in the southeastern piedmont of the Brooks Ridge, North Alaska (as deciphered from the airial photograph): (1) river channel, sandy and sandy–pebble bars; (2) vegetation represented by horsetails, herbaceous monocotyledons, and small willow shrubs; (3) vegetation represented by willow bushes; (4) predominantly coniferous (spruce) vegetation.

tation-free habitats at the early succession phases (Retallack and Dilcher, 1981a, 1986).

The dispersal of angiosperms was concurrent to that of advanced ferns and conifers, and they formed together the independent plant communities. The fact that "Mesophytic" elements were not found in association with "Cenophytic" plants in the OCSR localities of the Arman and Ayanka floras (which contain both elements, according to Samylina, 1974, and my own observations) suggests that not individual plants were replaced by more advanced ones, but the entire plant communities were involved in this process, and the advanced communities may locally coexist for some time with some older associations until they finally will replace the latter (Samylina, 1974).

The evolving Cretaceous angiosperms successfully competed with ferns and gymnosperms in many more stable habitats and developed their specialized ways of reproduction. The Cretaceous history of the AKSR platanoceans demonstrates their gradual penetration from unstable channel areas into more stable ones (Herman, 1994a). In the upper Albian–Lower Cenomanian deposits, platanocean occur only in fluvial sandstones, whereas in the Turonian–Coniacian deposits they were also found in finer grained, frequently carbonaceous sediments of the oxbow-lake and lacustrine facies (Koporulin and Voznesenskii, 1969), where they show the higher diversity (judging from the leaf morphology) and abundance.

However, it should be stressed that the invasion of angiosperms into vegetation of the early succession phases and respective colonization of the disturbed habitats did not necessarily result in replacing of the earlier vegetation: at present, coniferous forests are widespread over the vast territories. Angiosperms may acquire evolutionary advantages owing to some external processes (for example, climatic changes, intensive volcanic activity), which affected either the composition of plant associations of the mature succession phases (Krasiliv, 1992) or the relative competitiveness of angiosperms and other groups of higher plants.

The distribution of recent vegetation observed by me in the valley of an Alaska river flowing down the southwestern slope of the Brooks Range (Fig. 5) can give a rough idea of how the Cretaceous angiosperms replaced the conifer–fern vegetation along the river valleys of inner Asian areas. Along the river in Alaska, the vegetation corresponds to the forest-tundra, where forest massives are dominated by conifers. The coniferous plants grow in the bank areas undermined by the river. Newly formed sandy and sandy–pebble meander bars are occupied first by horsetails, herbaceous monocotyledons, and small willow shrubs, and then by willow bushes and balsam poplars. Conifers, which need more time to grow up in the new river bars, replace the willow bushes later at some distance from the bank.

STRATIGRAPHIC CONCLUSIONS

All the above said should be taken into consideration when elaborating the Albian–Upper non-marine Cretaceous stratigraphy for northeastern Asia.

The lower, middle, and basal upper Albian deposits constitute a single regional Buor-Kemyus phytostratigraphic horizon including the beds with the fossil flora of the Buor-Kemyus type (*Resheniya...*, 1978). The flora-bearing beds of this age are known in the VCSR and OCSR territories as well as in the neighboring areas of Asia and North America (Fig. 2), but they have not been detected in the AKSR. Subregions cannot be distinguished in northeastern Asia for this interval of the geological history.

It seems impossible to propose a single phytostratigraphy of the nonmarine Upper Cretaceous (including the upper Albian) of the entire region (Herman, 1993a, 1994b; Shchepetov, 1995a, 1995b; Belyi, 1997). I think it necessary to elaborate independent stratigraphical schemes for each of three subregions of northeastern Asia (Fig. 3, Table 2).

Some time ago, I substantiated the recognition of the upper Albian–Maastrichtian (Danian?) phytostratigraphic horizons of the AKSR (Herman, 1993a) corresponding to the evolutionary stages of subregional flora. Each horizon was taken as a combination of coeval formations or their parts; the main criteria for distinguishing a horizon were paleontological records. In my opinion, the middle–late Albian to Late Cretaceous floras of two other subregions of northeastern Asia have not been sufficiently studied to recognize there similar subdivisions. The VCSR and OCSR flora-bearing deposits of this age can be considered as "beds with corresponding flora." These beds are distinguished on the basis of fossil plant assemblages (taphofloras) of certain type (Table 2).

The flora-bearing deposits containing abundant and diverse remains of angiosperms can be dated via their correlation with reliably dated phytostratigraphic horizons of the AKSR and with successions of rich fossil floras of the NASR. The Cretaceous flora-bearing beds of these subregions are dated using their alternation with marine beds and/or according to the traceable replacement of one another along the strike.

It should be especially emphasized that the fossil plant successions of subregions mentioned above cannot be directly used for dating the Late Cretaceous floras of VCSR and OCSR, where they are dominated by ferns and gymnosperms. In addition, the ratio of relatively old to advanced elements is naturally variable in coeval floras of different types. Otherwise, provided the long existence of plant groups relict for the Late Cretaceous and the absence of invasion by young plant associations with pedominant angiosperms in the intercontinental OCSR and VCSR, the age of taphofloras of these subregions will be mistakenly considered older than it is.

The deposits containing remains of the Late Cretaceous "Mesophytic" plants should be dated using a combination of methods supplementing and controlling each other, for instance, it would be reasonable to look for lateral transitions from these deposits to those bearing taphofloras rich in angiosperm remains (which implies the abundance of plant remains and careful litpar-lit sampling) and to perform long-distance stratigraphic correlations using different biostratigraphic (especially palynological), lithological, and petrochemical methods, paleomagnetic data, and isotopic age determinations.

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Foraminifera Evolution and Dispersal in the Early Paleogene Postcrisis Epoch

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Abstract—Data on changes in the taxonomic composition of planktonic and benthic foraminifers across the Cretaceous–Paleogene boundary are presented. Extinction events in the highly specialized planktonic group were most remarkable. The new diversification stage began in the Danian starting with the appearance of primitive new-comers. The diversification went trough several phases before zonal boundaries became apparent. The extinction of benthic forms was of various character depending on the state of evolutionary development and on the facies confinement of their populations. Mass extinction occurred among the specialized forms that inhabited shallow shelf areas. The taxonomic diversity of more deep-water and genetically primitive groups decreased to a lesser extent. The revised stratigraphic and geographical ranges of some genera are presented along with examples illustrating the origin of new lineages and the progressive development of certain groups against the background of the degrading of others. The role of immigrant species in the regional assemblages is also elucidated.

Key words: *foraminifers, plankton, benthos, Cretaceous, Paleocene, extinction, regeneration, diversity.*

The Cretaceous–Paleogene boundary is one of the generally recognized crisis levels in the evolution of fauna. It marks the extinction of many faunal groups, including planktonic and benthic foraminifers. The revealed changes in fauna composition at this boundary led some researchers (*Cretaceous...*, 1979; Jauhri, 1996; and others) to the conclusion that the character of the extinction was gradual, not catastrophic as it is occasionally postulated. For instance, several stages of extinction of planktonic foraminifers were recorded in Tunisia (Keller, 1989). This event began at the end of the Late Cretaceous (29% of extinct species); about 26% of planktonic forms became extinct precisely at the Cretaceous–Paleogene boundary, and the rest of species gradually degraded up to extinction at the beginning of the Danian. Investigations in northern Africa (Blow, 1971), Spain (Smit, 1977), and other regions showed a gradual decrease in the abundance of planktonic genera and species in the Maastrichtian.

The lack of considerable changes in the benthic group of foraminifers was also recorded. According to Keller (1989), only 9% of benthic foraminifers did not pass to the Paleogene. However, most paleontologists suggested a more considerable extinction of benthos. In Poland, for example, 60 of 140 late Maastrichtian, mainly benthic species, i.e., 43%, have their last occurrence level at the Cretaceous–Paleogene boundary (see the paper by Pozaryska and Witwicka in *Cretaceous...*, 1979). Maslakova (1981) also reported on drastic changes in the foraminifer composition at this level. It should be noted, however, that different quantitative estimates of the first and last occurrence levels of generic and suprageneric taxa sometimes result from disagreements in foraminifer classification used by paleontologists. Nonetheless, my data from the northern Caucasus (Bugrova, 1996) also indicate a substantial extinction of fauna. The lowermost Paleocene sediments of Malyi Balkhan yielded only a few Maastrichtian forms.

The present comparison of planktonic and benthic foraminifera evolution in pre- and postcrisis epochs showed the following general sequence of events: extinction, survival, and subsequent diversification. However, the impact of the crisis on both ecologically different groups of foraminifers, and the rate of gaining the subsequent maximum diversity was dissimilar.

PLANKTONIC FORAMINIFERS

For the thermophilic planktonic globotruncanids, rotaliporids, and other forms, the Cretaceous–Paleogene boundary was actually the mass extinction level (Fig. 1), as it is manifested in the territory of the former USSR and other regions (Maslakova, 1981; Morozova, 1960; and others). In the Late Cretaceous, these groups progressively developed, displaying an increase in the taxa abundance and shell size, the appearance of new morphotypes, complication of morphology (including multiple apertures, plates, etc.) and ornamentation, a high density of populations, wide distribution ranges, and occurrence of numerous cosmopolitan species.

At the end of the Cretaceous, two large highly specialized superfamilies Globotruncanidea and Rotaliporidea became extinct, and this planktonic foraminiferal morphotype was lost. It should also be mentioned that, in the northern Caucasus, the plankton extinction occurred slightly below the Maastrichtian–Danian boundary (Bugrova, 1996).

The specialized forms coexisted with some morphologically primitive genera at the end of the Maastrichtian. Among the latter, there were *Globigerinella* forms with small smooth-walled tests, which continued to develop in the Paleogene, and genera of Heterohelicidae also characterized by small smooth or slightly spinose tests with globular chambers arranged in two or three rows. Among the 12 genera of this family known within the area of the former USSR (*Vvedenie...*, 1981), six genera of the most complicated morphology show their last occurrence level at the Cretaceous–Paleogene boundary. Shortly later, three more genera became extinct, and the single one that appeared in the Danian was unable to adapt to new environments. Two genera of this morphotype existed in the Paleocene for a time, but their species diversity was low, distribution areas were isolated, and the respective fossils are scarce.

Therefore, the planktonic foraminifer assemblages demonstrate the almost complete renewal at the Cretaceous–Paleogene boundary. The new order Globigerinida, representing another test morphotype, appeared. Their first rare specimens are recorded in the terminal Maastrichtian. Subsequently, they disappeared (the aplanktonic zone), and after a while, their progressive evolution began.

In the lowermost Danian, the diverse fauna of the *Parvulorugoglobigerina eugubina* Zone appeared and flourished. Its species had a thin and smooth multichambered test of a diminutive size (0.1–0.2 mm), which possessed features of adult specimens. The stage of their appearance and development was very short and not universally recorded. This fauna appeared in the Alpine basins and was recorded in the central Apennines, Syria, and western Kopetdag. Fauna of this type also appears to be characteristic of the Malyi Balkhan, Turkmenistan. Here, a thin bed overlying the sediments with the Maastrichtian fauna and underlying those with the Danian fossils hosts scarce planktonic shells, whereas other morphotypes of foraminifers are missing. In the central Ciscaucasia region (the Kuban River basin), the extremely small shells of planktonic species were found in brecciated limestones of the lowermost Danian.

Smooth-walled eoglobigerines appeared at the next stage of the plankton evolution (Morozova, 1960; Maslakova, 1981; see also the paper by Bang and Smit in *Cretaceous...*, 1979). Morozova described *Eoglobigerina trifolia, E. tetragona, E. pentagona, E. taurica, E. hemisphaera*, and others. This group of about 10 species is extremely impoverished as compared to the diverse and specialized fauna of the uppermost Cretaceous. The group was still of a low morphological and taxonomic diversity, but was almost "instantly" dispersed in both hemispheres (*Cretaceous...*, 1979, and later publications on deep-sea drilling data). Subsequently in the Danian, some of these species became extinct, and a new group of globigerines with a fine polygonal test ornamentation appeared and was, most likely, more progressive and replacing the smooth-walled eoglobigerines. New morphotypes that appeared were the smooth-walled globorotaloid tests of the genus *Planorotalites*, globoconuses characterized by a highly trochospiral test with fine-spinose walls and supplementary apertures, and primitive small spinose *Acarinina* forms represented by few species that thereafter gave rise to the outbreak of speciation. The first one in the Paleocene was the mass distribution of large multichambered forms with an open umbilicus, like *Acarinina inconstans* Subb., the index species of the corresponding zone, and *A. praecursoria* Moroz., as well as others. The *Globorotalia* species with a rather small compact test appeared at the same time. By the Middle Paleocene, planktonic foraminifers became larger, and diversification of globigerinids and other genera rapidly increased. The new Cenozoic stage of the plankton evolution began. The available data on its evolution enable an exact definition of the Cretaceous–Paleogene boundary in vast territories, and major events in the plankton development mark the boundaries of the zonal rank.

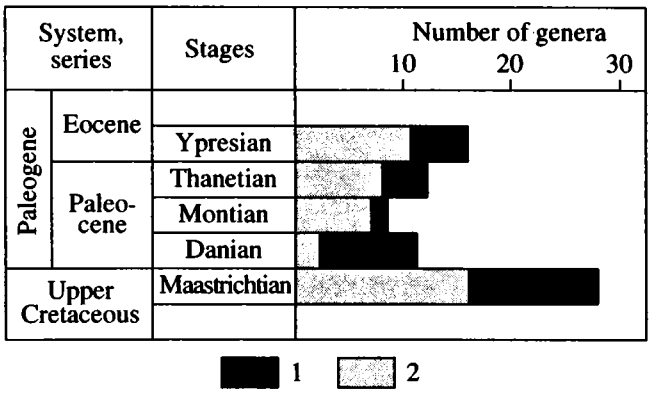

Fig. 1. Changes in generic diversity of planktonic foraminifers at the Paleocene boundaries (after Loeblich and Tappan, 1988, Fig. 22): (1) new genera; (2) genera of earlier origin.

At the beginning of the postcrisis evolutionary stage, plankton populations were numerous but isolated, and the species distribution areas are difficult to outline. Their evolution rate in the postcrisis epoch was most likely very high, and discriminative features of many taxa evolved unidirectionally. Thus, the definition of centers of the new taxa dispersal presents difficulties. However, I believe that the Alpine–Caucasus area of the Tethys was the possible cradle of globigerinids. During a short time of about 1.5 m.y. corresponding to the zones P1a and P1b, plankton dispersed

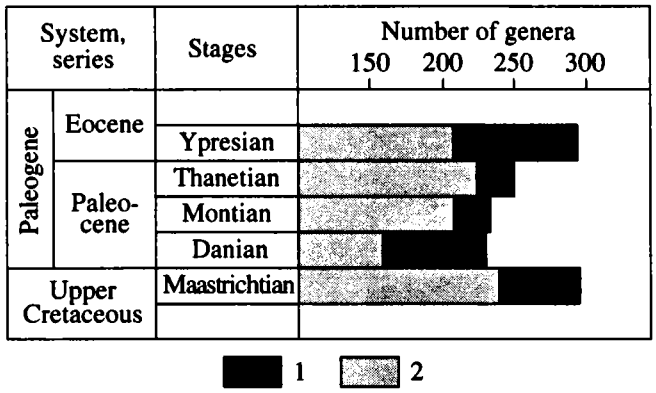

Fig. 2. Changes in generic diversity of benthic foraminifers at the Paleocene boundaries (after Tappan and Loeblich, 1988, Fig. 10): (1) new genera; (2) genera of earlier origin.

almost throughout the globe. In the basins of the tropical and subtropical zones, the speciation rate was high, so that it is possible now to distinguish 4 or 5 biostratigraphic zones based on certain plankton evolutionary stages. In the meridional direction, both the taxonomic and morphological diversity decreased, because not all immigrant species could survive in the cold-water basins.

BENTHIC FORAMINIFERS

Biotic events at the Cretaceous–Paleogene boundary affected the evolution of benthic foraminifers as well. According to American researchers (Loeblich and Tappan, 1988), 22 foraminifer families became extinct at that time, and 20 new ones appeared in the Danian (Fig. 2). Agglutinated specialized taxa represented 25% of the extinct families. Some morphotypes were subsequently lost. Representatives of the order Rotaliida were most diverse and abundant. The order diversity also decreased from 101 genera in the Maastrichtian to 89 in the Danian, though its representatives continue to exist nowadays.

Benthos variously responded to the crisis depending on the evolutionary development and facies confinement of its taxa. Some groups degraded slowly, others became rapidly and completely extinct. Extinction largely affected the highly specialized groups that inhabited shallow, most likely, to 50 m deep zones of carbonate accumulation in the Tethyan shelf seas (Bugrova, 1993, 1994). This can be illustrated by the evolution of orders Orbitoidida and Nummulitida.

Orbitoidids originated during the Santonian in the Alpine–Himalayan paleobiogeographical subregion (Grigoryan, 1986) and migrated to the western hemisphere at the beginning of the Campanian. In the Maastrichtian, they had a maximum taxonomic and morphological diversity, so that they were represented in both hemispheres by common genera and even species. In the eastern hemisphere, their extinction began in the Campanian. Initially, 2 genera and 6 species, and then 7 genera and 35 species, became extinct at the end of the Maastrichtian. The Cretaceous–Paleogene boundary was a critical level for this ecologically specialized group.

The appearance of *Discocyclina* forms at the beginning of the Paleocene marked the new evolutionary stage of Orbitoidida. The first representative of the genus had a small test and was primitive in morphology. This form is recorded in the lowermost Danian of the northern Caucasus (Bugrova, 1984) and, according to Grigoryan and Mrevlishvili, also occurs in younger lower Paleocene beds of Transcaucasia. By the Middle Paleocene, discocyclines were already represented by several species dispersed in the area from Spain to Syria and in America. Thus, after the extinction of specialized genera at the Cretaceous–Paleogene boundary, the new evolutionary stage of the order began and then terminated in the Eocene.

Along with orbitoidids, the first primitive nummulites appeared at the end of the Maastrichtian. Both groups were found in similar facies zones of Mangyshlak (Ashurov *et al.*, 1987), Tajikistan (Ashurov and Kreidenkov, 1973), northern Afghanistan (*Geologiya...*, 1980), and Armenia (Grigoryan, 1986). Rather scarce Danian nummulites occur in Crimea (Ashurov and Nemkov, 1991) and Pakistan (see the paper by Kureshy in *Cretaceous...*, 1979). This part of the Tethys area most likely was the procreation center of this new foraminiferal group that occupied the ecological niche left after the Cretaceous orbitoidids. Nummulites colonized the niche by the Middle Paleocene, but the distribution range of their three or five species was rather small and included Transcaucasia (Grigoryan, 1986; Mrevlishvili, 1984), northern Caucasus (Bugrova, 1984), and eastern Central Asia (Ashurov and Nemkov, 1978; Bugrova, 1991). However, at the end of the Paleocene, their assemblages still rather monotonous in composition already occupied southern Europe, northern and central Africa (Mali and Nigeria), Middle East, Transcaucasia, southern and eastern Central Asia, India, and Indonesia. They reached the maximum species diversity, namely, 120 species of 4 major genera, in the Middle Eocene and then gradually degraded.

A similar rapid development and degradation was characteristic of another specialized group of foraminifers, the order Rotaliida. In the Campanian, the first *Lockhartia* species appeared in the Caribbean. At the beginning of the Paleocene, they migrated to the eastern hemisphere, most likely via the Trans-Sahara Sea, as it is evident from paleontological records in Mali studied by V. Krasheninnikov and W. Berggren, and then became widespread in the marginal Tethyan areas (Butterlin and Foucard, 1989). The dispersal of *Lockhartia* and other thermophilic genera, namely, of *Kathina, Smoutina, Sakesaria, Daviesina, Ranikothalia*, and *Miscellanea*, along with orbitoidids and num-

mulites, was under the control of the climatic zoning. These groups inhabited the tropical and subtropical areas from the Caribbean to Afghanistan, Pakistan, and India. In northerly regions, *Lockhartia* forms are known from shallow, but still sufficiently warm basins of Western Europe (Hofker, 1966) and Central Asia, where they coexisted with nummulites and miliolids (Bugrova, 1966, 1991). When the listed rotaliid genera occupied the ecological niche of extinct orbitoidids in tropical seas, they rapidly adapted, progressively developed, and also soon degraded.

Inhabitants of deeper (down to 200 m) shelf areas and agglutinated foraminifers of a primitive morphology underwent the fewer changes, though some of their lineages that progressively developed before were interrupted.

In some instances, there was the gradual degradation of benthos at the last stage before the crisis. In the northern Caucasus, the Maastrichtian foraminifer assemblages included not less than 100 species, whereas their diversity at the end of this stage decreased to 20–25 species. Several opportunistic (in the terminology of Kauffman and Erwin, 1995) species of the genera *Cibicidoides, Stensioeina*, and *Osangularia* were dominant amid survived taxa, and others were relatively scarce (Bugrova, 1996).

Some Cretaceous forms (survivors) survived the crisis and inhabited a substantial territory at the beginning of the Paleocene, occasionally constituting a large fraction of an assemblage. Among them, there were *Plectina convergens* (Keller), *Bolivinoides delicatulus* (Jones), *Verneuilina kelleri* Moroz., *Stensioeina caucasica* (Subb.), *S. whitei* (Moroz.), *Gavelinella umbilicata* (Brotz.), *Marssonella oxycona* (Reuss), *Ataxophragmoides variabilis* (Orb.), and others. However, they did not give rise to new lineages.

Certain genera retaining their morphology lost the species diversity and subsequently became extinct after a certain period of development. For instance, the unidirectional development of morphological features without any divergence is established for the following taxa: *Coleites*, including the Paleocene *C. reticulosus* (Plumm.) and *C. unicus* Bugr. from the Lower Eocene of the Crimea; *Pyramidina*, including the Danian *P. curvisuturata* (Brotz.) and Zealandian *P. crassa* Brotz.; and *Karreria* (*K. fallax* Rzehak from the uppermost Cretaceous–Paleocene beds and *K. grata* Bugr. from the Lower Eocene of eastern Central Asia). The newcomers were either endemic, or had small and isolated distribution areas.

Other highly specialized and taxonomically diverse genera became extinct at the end of the Maastrichtian, but their lateral lineages continued to develop. The genus *Pseudogavelinella* Voloshina is an example. In the Cretaceous, it was represented by *"Anomalina" dainae* (Mjatl.), *A. infrasantonica* Balakhm., *A. costulata* Marie, *A. clementiana* (Orb.), and *A. pseudoexcolata* Kalinin (Vasilenko, 1954). The new species *P.*

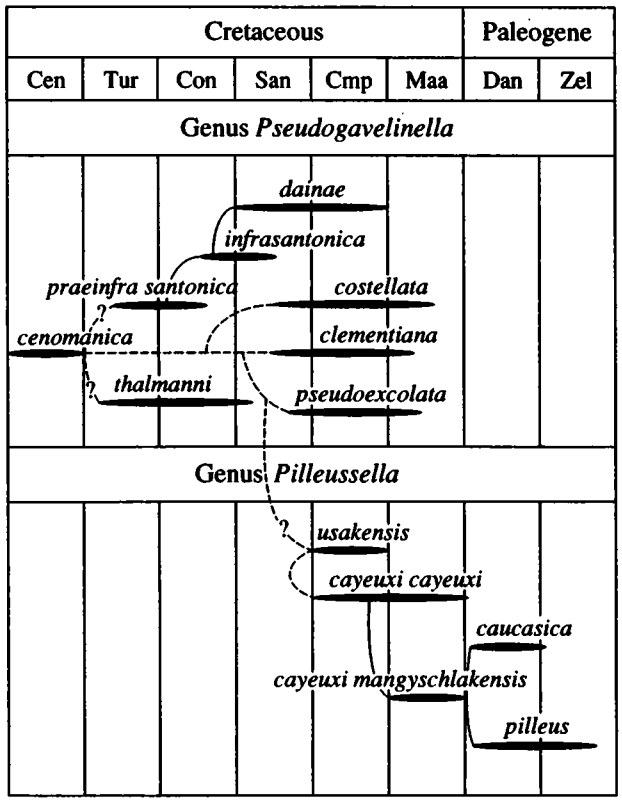

Fig. 3. Inferred genetic relations between *Pseudogavelinella* and *Pilleussella* species.

usakensis (Vass.) with a high central part of the spiral side appeared in the Campanian. This morphological feature is already characteristic of the genus *Pilleussella* J. Nikitina, whose representative *P. cayeuxi* (Lappar.) with subspecies *cayeuxi* and *mangyschlakensis* existed in the Maastrichtian and entered the Paleocene (Fig. 3). In the postcrisis time, this form coexisted with *Pilleussella caucasica* (Moroz.) of the Danian, and *P. velascoensis* (Cushm.) and *P. pilleus* (Vass.) of the Danian–Zealandian time. The two former species are representatives of deep-water assemblages from the northern Caucasus and Kopetdag, whereas *P. pilleus* is characteristic of a shallow-shelf, usually of the carbonate or carbonate-terrigenous facies in Mangyshlak and in the platform part of western Turkmenistan.

Some forms that appeared shortly before the crisis were subordinate in assemblages, after this event they became ancestors (or progenitor taxa) of new evolutionary lineages. For instance, species of *Falsoplanulina* and *Brotzenella* genera are of this type.

The ancestral species *Falsoplanulina ekblomi* (Brotz.) is known from the Maastrichtian. In the Paleocene, it produced two lineages, one of which terminated by the short-ranging specialized form *F. balakhmatovae* (Bugr.) at the end of the Early Eocene. Another lineage produced a number of transitional forms and evolved through the Eocene (Fig. 4).

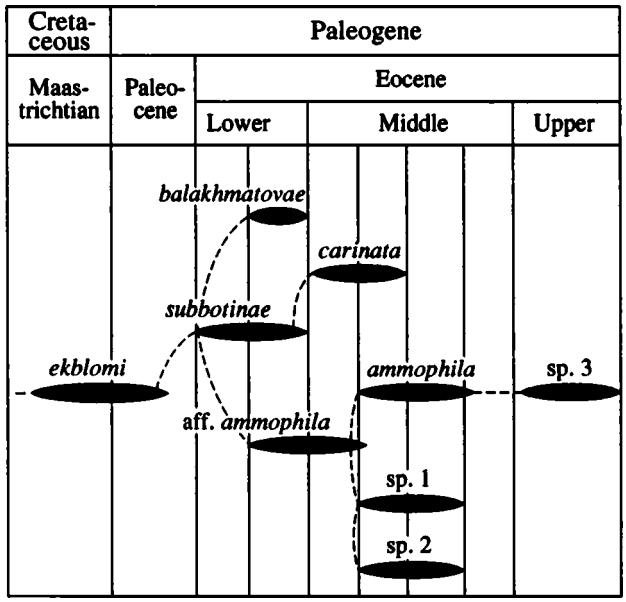

Fig. 4. Inferred genetic relations between the *Falsoplanulina* species.

The genus *Brotzenella* was divergent (Bugrova, 1983; Bugrova, 1996) at the early regeneration stages, and its survived species *B. praeacuta* (Vass.) gave rise to subspecies *praeacuta, elongata,* and *subsphaerica.* The subsequent evolution of the genus was similar to that in the Cretaceous displaying transmutation of small flattened forms into larger species with a great number of chambers in the outer whorl. The genus thriving in the Middle Eocene was characterized by a considerable species diversity and vast distribution areas of its highly dense populations. It is interesting to note a sudden appearance and disappearance of large tests of *Brotzenella* aff. *monterelensis* (Marie) sometimes referred to under the name "*taylorensis*" and similar to Cretaceous *B. dayi* (White), which had a lesser number of chambers. These species occur in sections of the northern Caucasus and Malyi Balkhan (Turkmenistan) within the *Globorotalia angulata* Zone. Their morphology is close to that of the Senonian species, and even the biometric methods are inadequate for their reliable distinction (Bugrova, 1978). This group can most likely be considered as the "lazarus taxon" having certain new features.

Abundant and diverse representatives of the order Nodosariida did not substantially change at the crisis level. Respective genera characteristic of the uppermost Cretaceous continued to evolve through the Paleocene. The changes affected mainly the specific level; however, this group apparently was in search for adaptive combination of former morphological features. In the Middle Paleocene (*Globorotalia angulata* and *G. conicotruncata* zones) of the northern Caucasus, Malyi Balkhan, and western Kopetdag, a very peculiar genus appeared and soon became extinct. Three generations of tests of its single species combine features of the genera *Lenticulina* (form B), *Planularia* (form A1), and *Palmula* (form A2), and, additionally, they all had a characteristic complicated morphology of the umbilicus resembling that of *Cribrolenticulina* forms. The *Lenticulina*-type tests of the generation B are considered as genetically conservative (Fursenko, 1978) and dominant in assemblages. Scarce tests of the generation A2 (divergent forms) were recorded only in western Kopetdag and seemingly represented the dead-end branch of the genus *Lenticulina.*

The planispiral coiling that appeared as a distinctive feature of the genus *Turkmeniella* (nom. correct. pro *Turkmenella* Bugrova, 1985) was a progressive feature in the evolution of lenticulinines (Bugrova, 1985). In the Paleocene, the genus species became widespread, especially *T. darwazaensis* (Balakhm.) characteristic of the Danian in western and central Turkmenistan and *T. klagshamnensis* (Brotz.) and *T. bzurae* (Pozaryska) typical of the Paleocene in southern Sweden, Poland, East European platform, and the Caspian lowland. The last of them was recorded in western Turkmenistan and Egypt (Speijer, 1994, pl. IX, Fig. 4). The genus flourished in the Middle Eocene.

The origination centers of both planktonic and benthic new-comers are usually difficult to establish. All they need is some time for migration. This time was insignificant for planktonic foraminifers, and in the Paleocene, they were dispersed almost throughout the globe and went through several stages of evolution. The distribution of benthos should meet with many geographical and ecologic barriers, but it also rapidly dispersed in the Paleocene because of the floating ability of its juvenile forms transported by surficial oceanic currents (Berggren and Aubert, 1975, Fig. 1) into shelf zones, whose arrangement was different than now (Barron *et al.*, 1981). Thus, the immigrant species played an important role in the regional postcrisis assemblages.

An example is the distribution of the Paleocene "Tethyan carbonate fauna" (in terms of Berggren, 1974) and of the Paleocene assemblages of the Midway and Velasco types (Berggren and Aubert, 1975, Fig. 1; Speijer, 1994; Petters, 1979; Morkhoven *et al.*, 1986; and others) in both hemispheres to the north and south of the equator.

The latter assemblages dispersed from the Caribbean, where they were originally described, over a vast territory, most likely, because of its transportation by equatorial currents and by the Gulf Stream, which headed to the European shore of the North Atlantic and reached, according to new findings of fauna, the southeastern coast of the Kara Sea. Fauna of the mentioned two types included dozens of cosmopolitan species and even more, if the unified taxonomic classification and nomenclature is used.

The Tethyan or Mediterranean fauna, though somewhat impoverished, was discovered in the Crimea by

E.K. Shutskaya along the Kacha River and by M.V. Yartseva in wells of the Bakhchisarai region, and also in the Tarkhankut Peninsula, eastern Central Asia, and northern Afghanistan (my determinations in collections of V.G. Morozova).

Many species of the Midway Shelf assemblage (depths of 50 to 200 m) are detected in different paleobiogeographical areas of the former USSR in the following regions: the East European plain, Scythian plate, Ciscaucasia, western Central Asia, the Caspian depression, Mangyshlak, and the Turgai depression. Species of the Velasco foraminifer assemblage (depth of dwelling over 200 m) are known from the Paleocene of the northern Caucasus and western Turkmenistan.

The influence of climatic zoning on the generic composition of benthos is particularly pronounced in foraminifer assemblages from Mali, Nigeria, southern Scandinavia, Poland, the East European plain, and southern and northern West Siberia. Their taxonomic and morphologic diversity decreases from the tropical regions towards the poles, whereas the significance of more primitive taxa correspondingly grows. The local conditions also affected the quantitative proportion between species of various genera, ecologic types of fauna, and endemic forms.

Therefore, analyzing the dynamic changes in taxonomic and morphological diversity of foraminifers, one can state that a sudden extinction of foraminiferal fauna at the Cretaceous–Paleogene boundary was catastrophic for certain taxonomic groups. This event was especially dramatic for planktonic foraminifers that inhabited surficial waters and mainly represented the taxa of a high evolutionary level. Few survived and more primitive morphological species were progenitors of new taxa and morphotypes.

At the same level, there was extinction of some benthic specialized taxa that inhabited the shallow-water settings. Their survived forms evolved for a time without considerable morphologic changes and then became extinct. Other forms experienced adaptive morphological transformations. The subsequent progressive development of foraminifers along various phyletic lines was rapid enough to result in the regeneration of the species diversity of foraminifers by the middle Paleocene.

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Palynostratigraphy and Dinocyst Zonation of the Middle Eocene–Lower Miocene Deposits at the Belaya River (Northern Caucasus)

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Abstract—The detailed biostratigraphic subdivision of a complete and well-exposed Paleogene–Miocene section in the Ciscaucasia is substantiated on the basis of organic-walled phytoplankton, spores, and pollen. The distinguished regional dinocyst zonation for the second half of the Eocene and for the Oligocene and lower Miocene shows the direct interlinking of zonal units and is correlated with the West and Central European zonal scales with the help of indicated microphytofossils.

Key words: *dinocysts, acritarchs, prasinophytes, phytoplankton, nannoplankton, spores, pollen, Eocene, Oligocene, Miocene, zonation, Ciscaucasia.*

INTRODUCTION

In contrast to the Kubanskoe section of Paleogene deposits in the Belaya River region, the Abadzekhskaya section (Fig. 1) has so far been ignored by stratigraphers because of the recorded hiatus in sedimentation, which spans the first half of the Eocene. However, its Bartonian–Priabonian and Oligocene intervals are exposed as well and completely as at the Kubanskoe site.

Korotkov (1935), who studied the Maikop Group at the Belaya River, was first to recognize the Abadzekhskaya Formation as an Oligocene sequence of brown carbonate-free clays with marl concretions and interbeds of dolomitic limestones. Grossgeim (1960) characterized the Belaya River sequence in detail and divided it into several formation in accordance with the Paleogene stratigraphic scale accepted for the northern Caucasus at the end of the 1950s. This researcher described the sequence lithology and substantiated its age on the basis of distribution of benthic foraminifers, mollusks, spores, and pollen. Unfortunately, the distribution of organic remains, phytofossils included, was characterized at that time (Grossgeim and Gladkova, 1953) not in a bed-by-bed manner, but at the formation rank only. More detailed characteristics of the sequence were presented in the 1970s by A.S. Stolyarov, whose data were used as a basis for a guidebook to the geological excursion of 1995 (*Excursion...*, 1995) for scientists involved into the IGCP projects no. 326 (*The Paleogene–Neogene Transition in the Northern Hemisphere*), no. 329 (*Neogene of the Paratethys*), and no. 343 (*Stratigraphic Correlation of the Peritethyan Epicratonic Basins*). An additional revision of the sequence was performed in 1994 and 1995 by S.V. Popov (Paleontological Institute, Russian Academy of Sciences), J. Krhovsky (Geological Institute, Czech Academy of Sciences), M.A. Akhmet'ev and E.P. Radionova (Geological Institute, Russian Academy of Sciences) during preliminary work before the excursion. These researchers also collected a series of samples for comprehensive paleontological study. Krhovsky studied nannoplankton and foraminifers, and I investigated phytoplankton, spores, and pollen from the same samples. Preliminary results of our study were published at the Conference of the IGCP Project no. 326 (Krhovsky *et al.*, 1995) and at the All-Russia Palynological Conference of 1996 (Zaporozhets, 1996).

In this paper, I summarized data on this and subsequent series of samples collected during the excursion by Yu.O. Gavrilov (Geological Institute, Russian Academy of Sciences) and myself from the Kuma and Belaya Glina formations of the Maikop Group. Several studied samples from the Morozkina Balka Formation are from the collection of 1996 by S.V. Popov (they are marked by letter index "p").

Separate intervals of the Maikop Group are known in the scientific community as formations with different names, including those not used in practice (e.g., the Abadzekhskaya Formation). In this connection, it should be mentioned that unit names are given here in accordance with nomenclature of *Excursion Gudebook* (1995). The respective stratigraphic column and sampling levels are shown in Fig. 2.

DESCRIPTION OF THE SEQUENCE

The sequence under discussion includes the following stratigraphic units: the Kuma (middle Eocene) and

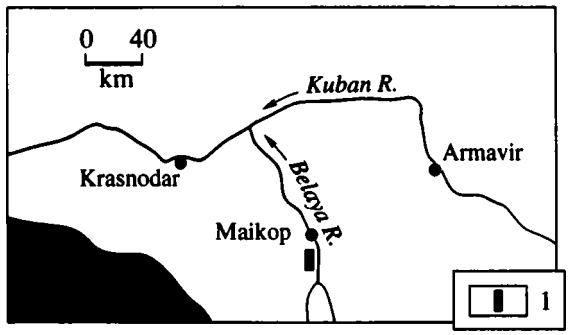

Fig. 1. Geographic position of the studied middle Eocene–lower Miocene section (1) in the Kuban–Belaya river system.

Belaya Glina (upper Eocene) formations and the Maikop Group. The lower part of the latter is divided into the Pshekha Formation, Ostracoda Beds, and Morozkina Balka Formation (lower to initial upper Oligocene). The middle subgroup includes the Batalpashi and Septarium formations (upper Oligocene and basal lower Miocene), whereas the upper one (lower Miocene) spans the Karadzhalga and overlying Ol'ginka formations hardly divisible on the basis of lithological criteria. The uppermost Ritsevka Formation of the Maikop Group appears to have been eroded in the pre-Tarkhanian time. Above this hiatus, there are Miocene deposits of the Tarkhanian Horizon, which have been sampled as well.

The *Kuma Formation* rests above the erosion surface of the Paleocene marine carbonates and includes beds of brown marl and marlaceous clay with marcasite concretions and fine detritus of plants and fish bones. The apparent thickness of the unit is 32 m. Near to the contact with the overlying Belaya Glina Formation, it is crowned with a transitional member (4.5 m) composed of alternating lighter and darker clayey marl beds characteristic, respectively, of the Belaya Glina and Kuma formations. The studied samples represent different levels of the Kuma Formation: its basal bed above the lower erosion surface (no. 3103) and levels of 3 m (no. 3105), 11 m (no. 3107), and 24 m (no. 3110) above the base. Samples 3111a and 3111d are from the base of the upper transitional member, and samples 3112a and 3112b represent the upper part of the latter.

The *Belaya Glina Formation* corresponds to the 20-m-thick unit of light-gray marlaceous clay and marl. At the top of the unit, there is a bed of greenish gray clay (1 m) bearing detritus of plants and fishes, and also the small thin-walled mollusks *Propeamussium fallax* and *Aturia ziczac* (determinations by I.A. Goncharova). Krhovsky attributed the main part of the Belaya Glina Formation to NP Zone 19 of the nannoplankton zonation by Martini (1971), and the *Propeamussium* Bed was correlated with NP Zone 20. The upper contact with the Maikop Group is sharp, but lacking the erosion marks. The Belaya Glina Formation is sampled at the base (Sample 3113) and at the level of 5–6 m below the top (Sample 95/119). Sample 95/87 is from the upper *Propeamussium* Bed.

The *Maikop Group* begins with the *Pshekha Formation* 25 m thick and divided into three members. The lower Member (I) of dark-gray to black calcareous clays (5–6 m) encloses remains of *Limacina (= Planorbella, Spiratella)*, fish imprints, Angiospermae leaves, and shoots of conifers (Taxodiaceae). The basal bed of the unit still corresponds to NP Zone 20. Samples 5 (with *Planorbella* forms) and 1 (with plant imprints) are from this member. The middle member (II) shows the rhythmical intercalation of gray calcareous and lighter clay beds (14–15 m), which are less calcareous or carbonate-free and bearing pteropod remains less frequent than below. The lower part of the member corresponds to NP Zone 21, whereas the upper one lacking pteropods is attributed to the higher NP Zone 22. Sample 21 is collected from the level of 4 m below the top of the member (10 m below the base of the Ostracoda Beds). The upper member (III) consists of dark-brown and black beds of carbonate-free clay (6 m) displaying inclusions of oxidized sulfides and fissures filled with jarosite and Fe-hydroxides. The upper boundary separates the unit from the overlying Ostracoda Beds of a lighter color. The last member is sampled at the levels of 3 m (no. 22), 1.5 m (no. 8), and 0.1 m (no. 9) below its top in order to detect changes in the phytoplankton composition by advancing toward the Ostracoda Beds. Nannoplankton is missing from this unit.

The *Ostracoda Beds* (= Polbinka Beds or synonymous regional horizon) represent the remarkable reference level of the sequence. They are composed of white or slightly darker marl bearing fossil ostracodes and nannoplankton remains of NP Zone 23 (0.3–0.5 m). The Ostracoda Beds are sampled at the base (Sample 24) and in their middle part (Sample 25).

The *Morozkina Balka* Formation lies over the erosion marks at the top of the Ostracoda Beds and is divided into two subformations. The lower subformation (40–50 m) consists of gray to brownish gray clay with jarosite efflorescence at the weathered surface. Its upper part encloses dolomitic nodules. Five samples (nos. 27, 29, 25p, 126, and 28p) characterize this interval of the sequence lacking nannoplankton. The upper subformation (35 m) is better exposed than other intervals of the Oligocene deposits at the Belaya River. In the center of the village of Abadzekhskaya, 0.5 km upstream of the suspension bridge, rocks of this subformation make up the whole series of cliffs at the right-hand side of the river. In contrast to underlying and overlying deposits, this unit is of a distinct flyschoid structure and shows an alternation of thin (2–5 cm) clayey and silty beds intercalated with rare beds of vitroclastic tuff. Four members (I–IV) can be recognized within the upper subformation. The lower Member I (8 m) represents the intercalation of clay beds, the brown carbonate-free ones with jarosite and Fe-

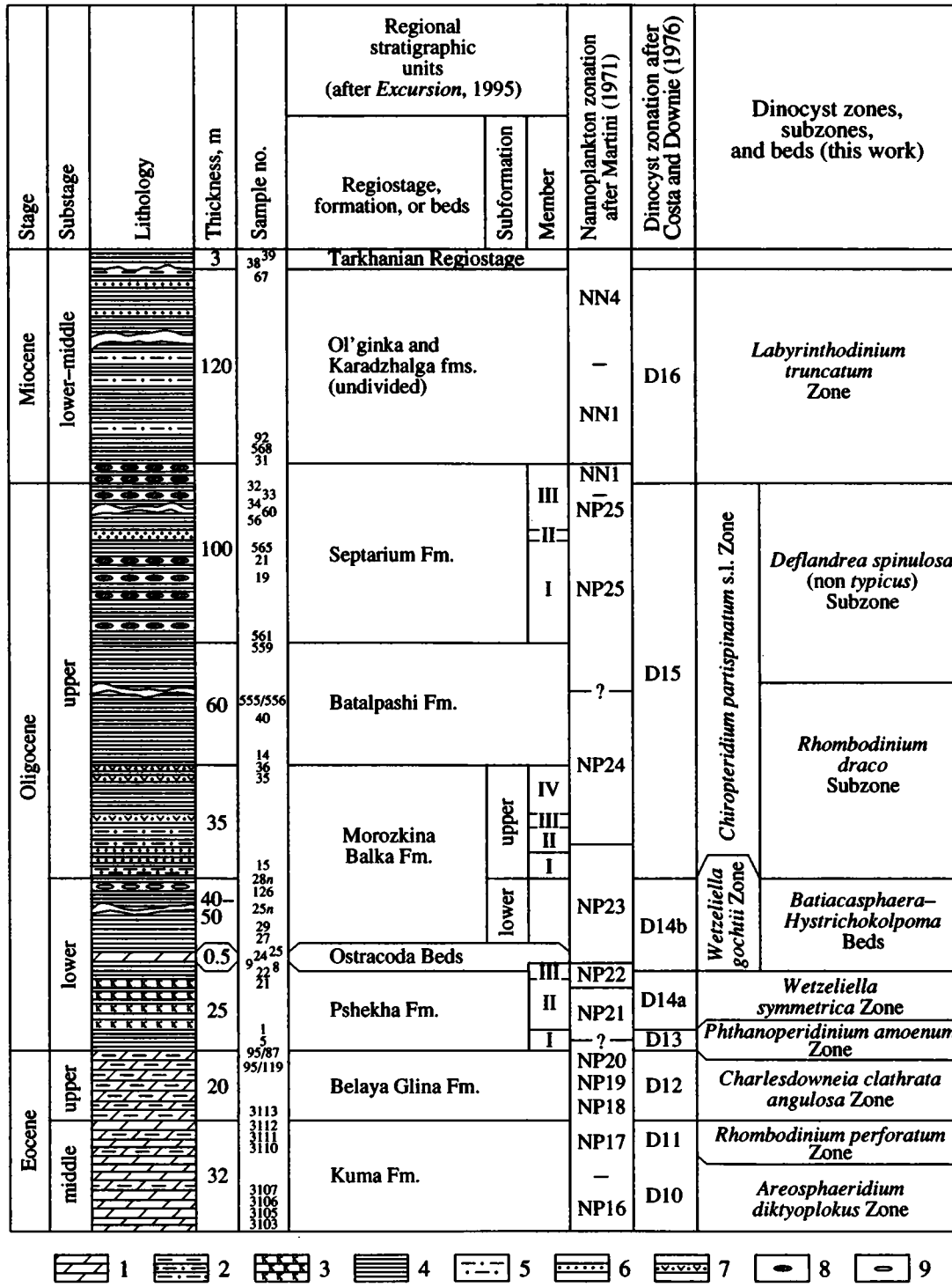

Fig. 2. Dinocyst zonation in middle Eocene–lower Miocene deposits at the Belaya River: (1) marl; (2) marlaceous clay; (3) calcareous clay; (4) slightly calcareous or pure clay; (5) silt; (6) sand; (7) tuff; (8) septaria; (9) dolomitic nodules.

hydroxides, and others, which are calcareous and colored gray. Member II (8 m) reveals five sedimentary cycles composed of calcareous to slightly calcareous clay beds bearing remains of batypelagic fishes and nannoplankton of NP zones 23 and 24 in the lower and upper parts of the member, respectively. Member III consists of carbonate-free to silty calcareous clay with jarosite (4 m) and encloses an ash interbed up to 1 m thick. The upper Member IV (12–15 m) consists of similar clay varieties though colored lighter than the

rocks below them. This unit also incorporates two interlayers of volcanic ash (3 and 2 cm thick). It yielded nannoplankton of the late Oligocene NP 24 Zone. The Morozkina Balka Formation is characterized by three samples: one (no. 15) from Member I and two (nos. 35 and 36) from Member IV, the latter collected at the level of 5–6 m above the thicker ash interlayer.

The *Batalpashi Formation* (60 m) consists of dark-brown clay enclosing jarosite and siderite segregations and marl nodules. The rocks are exposed at both sides of the Belaya River near the suspension bridge of the village of Abadzekhskaya and around the Molochka River mouth, the left tributary of the former. The lower boundary of the unit is distinct and divides the flyschoid rocks of the Morozkina Balka Formation and the more monotonous Batalpashi beds. The upper boundary separating the unit from the Septarium Formation is blurry and conventionally placed at the level where new flyschoid beds appear; septaria at this level are quite frequent and confined to the most calcareous rock varieties. Rocks of the unit are sampled in its lower (no. 14), middle (nos. 40 and 555/556), and upper (no. 559) parts.

The *Septarium Formation* is up to 100 m thick and includes two silty clay members separated by a bed of sand (0.9–1 m). The rocks are exposed at the northern slope of Mt. Lysaya, the left-hand side of the Belaya River. In its stratigraphic range, the Septarium Formation corresponds to the interval of the Alkun and Zelenchuk formations of the Kubanskoe section. Eastward of the Belaya River, in the Fars River valley (the right-hand tributary of the former), the relatively thin interlayer of sand grows in thickness up to 10 m. The interlayer rests on the basal conglomerate horizon and has been considered formerly as the individual Voskovaya Gora Formation (Grossgeim, 1960). At the Belaya River, the lower member (60 m) consists of carbonate-free to slightly calcareous silty clay bearing septaria. The upper member (40 m) of fine-grained rocks of the Septarium Formation is distinct because of the regular cyclic alternation of calcareous and carbonate-free clay beds. Nannoplankton of the formation is characteristic of the transitional NP 25–NN 1 interval. The contact with the overlying Karadzhalga Formation corresponds to the disappearance level of septaria. It is exposed near the Fyunt Creek mouth joining the Belaya River at the right-hand side downstream from the village of Abadzekhskaya.

The Septarium Formation is regularly sampled along its interval, and palynomorphs were counted in all 10 collected samples (nos. 561, 565, 19, 21, 56, 60, 32–34). This was necessary because of the disputable stratigraphic position of the Caucasian Regiostage attributed to the Miocene by Nosovsky and Bogdanovich (1979) and to the Oligocene by Andreeva-Grigorovich (1980).

The *Karadzhalga Formation* is the basal one in the upper Maikop Subgroup. The lower part of the formation (20 m) corresponds to monotonous beds of gray silty clay with jarosite. The contact between this and overlying Ol'ginka Formation is poorly exposed and assumed to be at the level, where sandy and calcareous rocks appear in the sequence. The Karadzhalga Formation is sampled in its lower part only (nos. 31, 568, and 92). The total thickness of the upper Maikop Subgroup is over 120 m. The uppermost member of the subgroup is sampled directly below the Tarkhanian deposits (no. 67). This level apparently corresponds to upper beds of the Ol'ginka Formation resting on the Karadzhalga Formation.

The *Tarkhanian Regiostage* is represented by deposits exposed in the steep right-hand wall of the river valley 2 km upstream from the Semikolennyi Creek mouth. In the lower part of the wall (10 m above the water level), the erosion surface of dark-gray carbonate-free clay of the Maikop Group is discordantly overlain by Tarkhanian beds of dark calcareous clay with a marl interbed (0.15 m) bearing remains of *Lentipecten denudatus* (determination by Goncharova). The Tarkhanian beds are 3 m thick and underlie the unit of sandy algal limestones (10 m) representing four persistent horizons of bioherms separated by thin interlayers of "smashed" clay. The early Tshokrakian age of the unit is inferred from molluscan fauna represented by characteristic species *Anadara bosphorana, Pteria micra, Ceritium orientale*, and others. Two samples (nos. 38 and 39) for this investigation are collected from clay interlayers separated by the marl bed.

PALYNOLOGICAL CHARACTERISTICS OF THE SEQUENCE

The Kuma Formation. All samples of the formation, those from the transitional member included, are saturated with amorphous organic matter and palynomorphs of higher plants. The less abundant phytoplankton of the unit is very diverse, although none of the distinguished taxa can be considered as dominant in abundance over others (table).

Samples of monotonous dark marl with fish remains and marcasite segregations (nos. 3103, 3105, 3107, and 3110b) yielded *Areosphaeridium diktyoplokus, A. arcuatum, Areosphaeridium* sp., *Adnatosphaeridium* sp., *Achomosphaera* sp., *Achilleodinium* sp., *Calligodinium* sp., *Cerebrocysta* sp., *Cerodinium diebelii, Cerodinium* sp., *Chytroeisphaeridia* sp., *Cleistosphaeridium* sp., *Cordosphaeridium* spp. (*C. exilimurum* included), *Cribroperidinium tenuitabulatum, Deflandrea* spp., *Diphyes* sp., *Horologinella* sp., *Hystrichosphaeropsis* sp., *Impagidinium* sp., *Impletosphaeridium* sp., *Lejeunecysta globosa, Manumiella* sp., *Microdinium* sp., *Operculodinium* sp., *Pentadinium* sp., *Phthanoperidinium sp., Rottnestia borussica, Selenopemphix nephroides, Spiniferites* sp., *Tectatodinium* sp., *Wetzeliella astra*, and *W. echinulata.* These forms associate with prasinophytes and acritarchs *Paucilobimorpha granuligera, Crassosphaera* sp., *Cymatiosphaera* sp., large

Leiosphaeridia sp., and *Ptanospermella* sp. In each sample, about 20% of specimens turned out to be fragmentary and indeterminable.

The middle Eocene (Bartonian) age of the unit is supported by data on foraminifers and nannoplankton from other sites in the northern Caucasus (Krasheninnikov and Muzylev, 1975). The organic-walled phytoplankton from the Kuma Formation is also comparable in composition with the Bartonian assemblages from West Europe and southern areas of the former USSR, where they characterize the *Areosphaeridium diktyoplokus* and *Rhombodinium perforatum* zones. The revealed distribution of index species shows that the greater part of the formation below the transitional member corresponds to the former lower zone.

The analyzed spore-and-pollen assemblage includes specimens of *Pinus* spp., *Betula* sp., *Myrica* sp., *Ilex* sp., *Rhus* sp., *Nyssa* sp., and formal pollen taxa, *Tricolporopollenites cingulum* included.

The dinocyst assemblage from the transitional member is of similar composition in general. *Areosphaeridium* and *Microdinium* species still are dominant here, but *A. diktyoplokus* is replaced for *A. arcuatum*, and new-comers are represented by *Rhombodinium* forms (*R. draco, R. rhomboideum*), *Pterodinium* sp., *Microdinium reticulatum, Melitasphaeridium asterium, Thalassiphora pelagica, Cyclopsiella* sp., *Dapsilidinium* sp., *Systematophora* sp., *Heteraulacacysta* spp. (*H. leptalea, H. pustulata*), *Batiacasphaera sphaerica*, and *Palaeocystodinium golzowense*. Green algae also remain almost unchanged in composition. The comparison of sample pairs, which were collected 2 cm apart from darker (nos. 3111a and 3112b) and lighter (3111d and 3112a) rock varieties of the basal and middle levels of the transitional member, demonstrated that phytoplankton is more diverse in the light colored rocks (more that 30 and 40 taxa in samples 3111d and 3112a versus 11 and 15 taxa in samples 3111a and 3112b).

Palynospectra from the transitional member show predominance of angiosperm taxa, especially of evergreen Fagaceae (*Quercus gracilis, Q. graciliformis, Q. conferta, Castanopsis* sp.). The less abundant evergreen taxa are *Magnolia* sp., *Engelhardtia* sp., *Rhus* sp., *Platycarya* sp., *Ilex* sp., Araliaceae, *Nyssa* sp., and some others. Pine pollen is dominant in the group of conifers also incorporating rare pollen grains of *Ephedra, Podocarpus*, and Taxodiaceae. In general, palynomorphs of higher plants are much more abundant in the dark marl beds characterizing the anoxic environments of sedimentation than in the white pelagic marlaceous clay.

The Kuma interval in the studied and Kubanskoe sections is characterized by the following taxa in common: *Deflandrea* spp., *Rhombodinium draco, Areosphaeridium arcuatum, Impagidinium* sp., *Wetzeliella echinulata, Palaeocystodinium golzowense, Microdinium* sp., *Hystrichokolpoma* sp., *Horologinella* sp., *Pentadinium* sp., and *Manumiella* sp. In addition, *Dracodinium condylos, Samlandia reticulifera, Apteodinium* sp., *Ceratiopsis* sp., *Fromea* sp., *Melitasphaeridium* sp., *Pentadinium lophophorum*, and *Charlesdowniea coleothrypta* appear in the Kubanskoe section, where the transitional member shows predominance of *Chytroeisphaeridia* forms (30%) over other taxa and presence of *Trigonopyxidia fiscellata* (8%) and *Thalassiphora pelagica* (more than 5%).

The Belaya Glina Formation. The diverse phytoplankton of the formation includes dozens of taxa and represents up to 90% and more of the total amount of palynomorphs. Dwellers of the outer shelf zone, such as *Achomosphaera* spp. (*A. sagena* included), *Impagidinium* sp., *Spiniferites* spp., and *Systematophora* sp., are dominant at this level. Characteristic taxa of the late Eocene are represented by *Areosphaeridium arcuatum, Cordosphaeridium funiculatum, Rottnestia borussica*, and *Tytthodiscus beloglinensis*. All the listed taxa coexist with zonal index species of the *Charlesdowniea clathrata angulosa* Zone of the West European scale. Rare specimens missing in the overlying Oligocene deposits belong to *Achilleodinium* sp., *Pterodinium* sp., *Corrudinium* sp., *Lophocysta* sp., *Cannosphaeropsis* sp., *Dinopterigium* sp., *Wilsonidium* sp., *Fibrocysta* sp., *Turbiosphaera* sp., and *Microdinium* sp.

In a sample from the formation base, phytoplankton represents 89% of all palynomorphs. The summary abundance of *Impagidinium* spp. (12%), *Areosphaeridium* spp. (mainly *A. arcuatum*; 9%), and *Corrudinium incompositum* (9%) is 30%. *Tectatodinium* sp. (7%), *Thalassiphora* sp. (7%), and *Batiacasphaera* spp. (*B. hirsuta* and *B, sphaerica* included; 5% in sum) are next in abundance. *Operculodinium* spp. (*O. placitum* and *O. tiara* included) and *Hystrichokolpoma cinctum* represent 3% each. Abundance values slightly higher than 1% are characteristic of *Rhombodinium* spp. (*Rh. rhomboideum* and *Rh. pustulatum* included), *Pentadinium* spp. (*P. taenigerum* amid them), *Spiniferites* spp., *Achomosphaera* sp., *Microdinium* sp., *Deflandrea* spp. (*D. phosphoritica* included), *Melitasphaeridium asterium*, and *Cribroperidinium tenuitabulatum*. Single specimens are identified as *Pyxidiella* sp., *Samlandia* sp., *Diphyes colligerum, Gochtodinium* sp., *Apteodinium* sp., *Phelodinium* sp., *Wetzeliella echinulata, Rottnestia borussica, Lentinia* sp., *Lejeunecysta globosa, Cerebrocysta* sp., *Fromea* sp., *Achilleodinium* sp., *Areoligera* sp., *Cyclopsiella* sp., and *Tytthodiscus beloglinensis*. Taxa occurring in and above the unit are *Lingulodinium* sp., *Selenopemphix nephroides, Homotryblium plectilum*, and some others.

Sample 95/119 yielded the dominant *Impagidinium* spp., *Systematophora* spp., *Achomosphaera* sp., and *Spiniferites* spp. in association with rather frequent specimens of *Pterodinium* sp., *Operculodinium* sp., *Areosphaeridium arcuatum*, and *Cordosphaeridium* sp. The total number of taxa is over 40 and grows up to 55 in the *Propeamussium fallax* Beds at the top of the for-

Table. Distribution patterns of organic-walled phytoplankton in the middle Eocene–lower Miocene section at the Belaya River (western Ciscaucasia)

Species	Eocene				Oligocene								Miocene	
	middle		upper		lower					upper			lower	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Areosphaeridium diktyoplokus	m	+	o											
Areosphaeridium arcuatum	x	m	+	+	o	o	o							
Cerodinium diebelii	o													
Cordosphaeridium exilimurum	o								o	o				
Cribroperidinium tenuitabulatum	o		+		o		o							
Lejeunecysta globosa	o		o							o			o	
Rottnestia borussica	+	o	x											
Selenopemphix nephroides	o		o				o			o	+		o	o
Wetzeliella astra	o													
Wetzeliella echinulata	o	o												
Achomosphaera sp.	o			x		o			o	o			o	o
Achilleodinium sp.	o		o											
Calligodinium sp.	o													
Cerebrocysta sp.	o		o							o				
Chytroeisphaeridia sp.	o					o	o		x	o	o			
Cleistosphaeridium sp.	+					o	o	o		x	o	o	o	o
Diphyes sp.	o													
Horologinella sp.	o				o	o	o	o	o	o				
Hystrichosphaeropsis sp.	o									o			o	
Impagidinium sp.	x	+	x	x				o	o	o				
Impletosphaeridium sp.	o													
Manumiella sp.	o													
Paucilobimorpha granuligera	+													
Cymatiosphaera sp.	+									o		+	+	
Leiosphaeridia sp.	+									x		+	m	
Tectatodinium sp.	o		x			o	o	o	o	o				
Rhombodinium perforatum		o												
Rhombodinium draco		o							+	o				
Rhombodinium rhomboideum		o	+											
Microdinium reticulatum		m												
Melitasphaeridium asterium		o	+			o	o				o			o
Thalassiphora pelagica		o	+			o				o			o	
Heteraulacacysta leptalea		o												
Heteraulacacysta pustulata		o												
Batiacasphaera sphaerica		o	+						m	o				o
Palaeocystodinium golzowense		o	o				o			x				
Cyclopsiella sp.		o	o											
Pyxidiella sp.			o			o							o	
Samlandia sp.			o											
Gochtodinium sp.			o											
Phelodinium sp.			o						o	o				
Lentinia sp.			o											
Fromea sp.			o										o	
Areoligera sp.			o											
Tytthodiscus beloglinensis			o	o										
Achomosphaera sagena			x											
Cordosphaeridium funiculatum			o											

Table. (Contd.)

Species	Eocene				Oligocene								Miocene	
	middle		upper		lower					upper			lower	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Charlesdowniea clathrata angulosa			o											
Corrudinium incompositum			+						o					
Hystrichokolpoma cinctum			+											
Batiacasphaera hirsuta			+	o						o				o
Operculodinium tiara			+											
Operculodinium placitum			+											
Rhombodinium pustulosum			+											
Pentadinium taenigerum			+											
Deflandrea phoshoritica			+		o		o			x				
Diphyes colligerum			o											
Homotryblium plectilum			o											
Phthanoperidinium amoenum					o	o	+	o						
Homotryblium floripes					o		o							
?*Cyclonephelium* sp.					o									
Gonyaulacysta sp.					o									
Lejeunecysta hyalina					o	x	+		o	x				
Hystrichosphaeridium sp.						o	o							
Areosphaeridium pectiniforme						o	o	o	o					
Adnatosphaeridium multispinosum						o								
Dapsilidinium simplex						o							o	
Pentadinium goniferum						o								o
Homotryblium tenuispinosum						o				x	o	+		o
Homotryblium aculeatum						o								
Wetzeliella symmetrica						o	o			x				
Bitectatodinium sp.						o			o					
Selenopemphix coronata						x								
Phthanoperidinium comatum						o				o				
Phthanoperidinium echinatum						o								
Systematophora placacantha						o								o
Batiacasphaera baculata							o	x	m	o				
Cordosphaeridium gracile							o							
Kallosphaeridium sp.							o	o	o					
Dapsilidinium pastielsii							o							
Membranophoridium aspinatum							o				o			
Wetzeliella gochtii							o	o	m					
Gerdiocysta sp.							o	o	o					
Hystrichokolpoma rigaudiae							o	o	+				o	
Hystrichokolpoma salacium							o	o	+		o		o	
Hystrichokolpoma spp. (Ostracoda Beds)							o	m	m	o				
Spiniferites elongatus								+						
Spiniferites cf. *mirabilis*								+						
Batiacasphaera micropapillata								x	m					
Leberidocysta sp.								o						
Stoverocysta sp.								o						
Pterodinium cingulatum								o						
Paralecaniella indentata								+	o					
Chiropteridium partispinatum								o	+	m	o	o		
Apteodinium emslandense									o	x				
Cribroperidinium cf. *guiseppei*									o					
Distatodinium ellipticum									o					

Table. (Contd.)

Species	Eocene				Oligocene								Miocene	
	middle		upper		lower					upper			lower	
	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Deflandrea spinulosa (typicus)									o	x	+			
cf. *Glaphyrocysta inculta*									o					
Cleistosphaeridium ?insolitum									o					
Pentadinium laticinctum									o	o	o			
Prolixosphaeridium sp.									o					
Chiropteridium mespilanum										x				
Dapsilidinium colligerum										o				
Paucilobimorpha triradiata										o				
Lejeunecysta paratenella										x			o	o
Operculodinium microtrianum										o				
Lingulodinium machaerophorum										o				o
Spiniferites pseudofurcatus										o				
Spiniferites ramosus										o				x
Wetzeliella articulata										o				
Deflandrea andromensis										o				
Deflandrea heterophlycta										x				
Apteodinium maculatum										o			o	
Cometodinium sp.										o				
Cyclopsiella vieta											o			
Heteraulacacysta campanula											o			
Gerlachodinium aechmophorum											o			
Deflandrea spinulosa (small cysts)											+	m	o	
Labyrinthodinium truncatum													o	
Homotryblium oceanicum													o	
Pyxidiella scabrosa													o	
Spiniferites bentorii														+
Operculodinium israelianum														+
Operculodinium centrocarpum														m
Concentricysta sp.														o
Tectatodinium psilatum														o
Tectatodinium pellitum														o
Tuberculodinium vancampoae														o
Sumatrodinium sp.														o
Riculacysta sp.														o
Distatodinium paradoxum														o
Hystrichosphaeropsis obscura														o
Cordosphaeridium inodes														o
Lejeunecysta oliva														o
Lejeunecysta sabrina														o
Trinovantedinium sp.														o
Apteodinium piasecki														o

Unit numbers: (1) Kuma Fm. (main part); (2) Kuma Fm. (beds transitional to the overlying unit); (3) Belaya Glina Fm. (lower part); (4) Belaya Glina Fm. (upper part); (5) Pshekha Fm., Member I (*Planorbella* Beds); (6) Pshekh Fm., Member II; (7) Pshekh Fm., Member III; (8) Ostracoda Beds; (9) Morozkina Balka Fm. (lower subformation); (10) Morozkina Balka Fm. (upper subformation); (11) Batalpashi Fm.; (12) Septarium Fm. (main part); (13) Septarium Fm. (terminal part), i.e., Ol'ginka and Karadzhalga fms. (undivided); (14) Tarkhanian Regiostage. Abundance of taxa (relative % in assemblage): (m) more than 10%; (x) 5 to 10%; (+) 1 to 5%; (o) less than 1%.

mation. In these beds, however, none of the taxa is more abundant than 5%, though *Deflandrea* spp., *Impagidinium* spp., *Operculodinium* spp., *Spiniferites* spp., and *Rottnestia borussica* are more frequent here than others. The high abundance of green algae and acritarchs, which was established by Savitskaya (1996) at the top of the Belaya Glina Formation in Ukraine, is untypical of the section under consideration.

Spores and pollen from the Belaya Glina Formation are less diverse than in underlying and overlying deposits, because sediments of the unit have been accumulated in a distal seaward zone. Dominant specimens of *Pinus* spp. coexist at this level with the diverse tricolpate pollen, and with that of *Ephedra* sp., *Podocarpus* sp., Juglandaceae, Fagaceae (mainly evergreen *Quercus graciliformis* and *Q. conferta*), *Rhus* sp., Moraceae, Hamamelidaceae, *Myrica* sp., and *Ilex* sp. A higher

diversity is characteristic of pollen from upper beds of the formation, where Taxodiaceae appear in the assemblage together with less frequent *Picea* sp., *Tsuga* sp., *Keteleeria* sp., *Cedrus* sp., *Ephedra* spp. (*E. fusiformis* and *E. claricristata*), and *Liqudambar* sp. In this situation, phytoplankton is somewhat less abundant than pollen representing mainly conifers, especially the *Pinus* group.

The Pshekha Formation. Phytoplankton was studied in all three members of the formation. In comparison with the Belaya Glina Formation, its abundance and diversity considerably decrease already in the basal *Planorbella* Beds (Member I), and many taxa characteristic of the underlying unit (*Rottnestia borussica, Tytthodiscus beloglinensis, Areosphaeridium dyktioplokus*, and others) are missing from this level. An important new-comer is *Phthanoperidinium amoenum*, the index species of the lower Oligocene zone of the West European scale. At the Belaya River, it appears earlier in the sequence than *Wetzeliella symmetrica* and *W. gochtii*. In Sample 5 collected at the level of 2–2.5 m above the base of the Maikop Group, dominant taxa are *Homotryblium* spp. (*H. floripes* included), *Cyclonephelium* sp., *Deflandrea* spp. (*D. phosphoritica* amid them), *Gonyaulacacysta* sp., *Horologinella* sp., *Hystrichosphaeridia* sp., *Lejeunecysta hyalina, Cribroperidinium tenuitabulatum*, and *Pyxidiella* sp. As compared to the Beds, where phytoplankton represents at least 40–45% of palynomorphs, its abundance in the *Planorbella* Beds is 5% only.

Spores and pollen from the *Planorbella* Beds are diverse. Pollen of conifers is most abundant (over 70%), and proportion between pine and Taxodiaceae pollen is 3 : 1. *Picea* sp., *Tsuga* sp., and *Ephedra* sp. are represented by single pollen grains. Pollen of Betulaceae (especially of *Carpinus* spp.) is dominant in the Order Betulales, whereas that of *Carya* sp. and deciduous oaks prevails amid specimens of Juglandaceae and Fagaceae, respectively. The abundance of *Rhus* sp. pollen is abnormally high (over 10%). Other taxa of the assemblage are Hamamelidaceae (*Hamamelis* sp. and *Corylopsis* sp.), Moraceae, Myricaceae, and Araliaceae.

In addition, the well-preserved leaf remains of Lauraceae (*Cinnamomophyllum*), Myricaceae, *Dryophyllum* sp., and shoots of *Cryptomeria* sp. and *Taxodium* sp. (determinations by Akhmet'ev) occur in the *Planorbella* Beds.

Member II (beds of calcareous to slightly calcareous clay) was sampled in its upper part (Sample 21, 10 m below the Ostracoda Beds). Phytoplankton is not abundant at this level. Its dominant forms are *Lejeunecysta* spp. (*L. hyalina* included) and *Selenopemphix* spp. (*S. nephroides* and *S. coronata* amid them). Species of the genus *Phthanoperidinium* are quite diverse, but scarce (*P. comatum, P. amoenum*, and *P. echinatum*). The member corresponds to the first occurrence level of *W. symmetrica*, the index species of the synonymous dinocyst zone of the Rupelian in the West European scale. *Areosphaeridium* forms are represented by *A. pectiniforme* and *A. arcuatum*, the latter of which usually resides below the Eocene–Oligocene boundary in the northern part of the region and survives up to the early Oligocene in the southern warmer part of the respective paleobasin. *Adnatosphaeridium multispinosum* was also encountered in the member, where newcomers missing from underlying beds are represented by *Cleistosphaeridium* sp., *Dapsilidinium simplex, Hystrichosphaeridium* sp., and *Pentadinium goniferum*. Single specimens correspond to the following taxa: *Batiacasphaera* sp., *Bitectatodinium* sp., *Apteodinium* sp., *Chytroeisphaeridia* sp., *Melitasphaeridium asterium, Palaeocystodinium golzowense, Systematophora placacantha, Thalassiphora pelagica*, and *Tectatodinium* sp. Among prasinophytes and acritarchs, there were detected *Micrhystridium* sp. and *Ptanospermella* sp.

The spore–pollen assemblage of the member includes pollen of conifers as the dominant component (more than 75%; Taxodiaceae/Pinaceae ratio 3.5 : 1). In contrast to the assemblage of *Planorbella* Beds, this one is more rich in Fagaceae, especially in pollen of evergreen oaks and *Castanopsis*. Other thermophilic taxa, such as *Corylopsis* sp., *Hamamelis* sp., *Ilex* sp., Sapotaceae, *Aralia* sp., and *Engelhardtia* are more diverse in this member than below.

Member III is impoverished in phytoplankton, whose endemic taxa characteristic of the Ostracoda Beds appear for the first time in the uppermost deposits of the underlying member.

Deflandrea spp., *Hystrichosphaeridium* sp., *Phthanoperidinium* spp. (mainly *P. amoenum*), *Lejeunecysta hyalina*, and *Operculodinium* spp. appear to be more abundant than other forms of the member. Single specimens represent here a wide spectrum of taxa: *Achomosphaera* sp., *Areosphaeridium arcuatum, A. pectiniforme, Apteodinium* sp., *Batiacasphaera* spp. (*B. baculata* included), *Cleistosphaeridium* sp., *Cordosphaeridium* spp. (*C. gracile* amid them), *Chytroeisphaeridia* sp., *Cribroperidinium tenuitabulatum, Deflandrea phosphoritica, Dapsilidinium pasteilsii, Kallosphaeridium* sp., *Lingulodinium* sp., *Melitasphaeridium asterium, Palaeocystodinium golzowense, Selenopemphix nephroides, Systematophora* sp., *Tectatodinium* sp., and *Wetzeliella symmetrica*. Directly below the contact with the Ostracoda Beds, there appear *Wetzeliella gochtii, Membranophoridium aspinatum, Gerdiocysta* sp., *Horologinella* sp., *Homotryblium* spp. (*H. floripes* included), and several thin-walled *Hystrichokolpoma* morphotypes widespread in the Ostracoda marl outcrops at the Belaya and Kuban rivers.

The spore–pollen assemblage of the member again shows predominance of coniferous pollen (Taxodiaceae/Pinaceae ratio 9 : 1). Fagaceae and Juglandaceae forms are of a similar abundance. Pollen of Betulaceae

decreased in abundance at this level, and that of *Comptonia* sp., *Rhus* sp., *Nyssa* sp., and *Fraxinus* sp. occurs as single grains.

The Ostracoda Beds. These beds mark significant changes in the distribution of dinocyst taxa and the appearance of new forms in response to the freshening of the paleobasin. First of all, this event affected representatives of genera *Hystrichokolpoma* and *Batiacasphaera* (Akhmet'ev and Zaporozhets, 1996) representing in sum 85–90% of the phytoplankton assemblage. In addition to usual *H. rigaudiae* and *H. salacium*, the assemblage includes the thin-walled *Hystrichokolpoma* forms representing seven or eight individual morphotypes. They all coexist with three *Batiacasphaera* species typical of the Oligocene: *B. sphaerica* (dominant), *B. baculata*, and *B. micropapillata.* Other taxa persistently occurring in the Ostracoda Beds are *Spiniferites* spp. (*S. elongatus* and *S.* cf. *mirabilis* are especially characteristic, and, in contrast to other representatives of the genus, they appear first in the brackish-water but not in the open-sea sedimentary facies), *Horologinella* sp., *Gerdiocysta* sp., cf. *Leberidocysta* sp., *Stoverocysta* sp., *Pterodinium* spp. (*P. cingulatum* sbsp. *reticulatum* included), *Impagidinium* sp., *Tectatodinium* sp., *Cleistosphaeridium* sp., *Oligosphaeridium* sp., *Corrudinium incompositum, Paralecaniella indentata, Distatodinium* sp., cf. *Caligodinium* sp., *Areosphaeridium pectiniforme, Kallosphaeridium* sp., *Phthanoperidinium amoenum, Wetzeliella gochtii*, and cysts of the *Glaphyrocysta–Cyclonephelium–Adnathosphaeridium* group. The beds also correspond to the first occurrence level of *Chiropteridium partispinatum.*

In contrast to sections at the Belaya River, those in the Ciscaucasia, where they correspond to marginal deposits of the Solenovian basin of the East European platform (the northern Ergeni area), and also in the Turan platform, show that phytoplankton of the Ostracoda Beds is enriched in *Micrhystridium* sp. and green algae. This apparently is a consequence of more significant freshening in the peripheral parts of the basin in response to the fresh water influx from the adjacent land areas.

On the basis of abnormal composition of phytoplankton, the Ostracoda Beds can be considered as corresponding to the local *Batiacasphaera–Hystrichokolpoma* Subzone of the *Wetzeliella gochtii* Zone.

The beds also show changes in the pollen spectrum of conifers. Pinaceae, mainly *Pinus* spp., slightly prevail here over Taxodiaceae. Juglandaceae (*Carya* sp., *Juglans* sp., *Engelhardtia* sp.) along with Fagaceae, including the evergreen oaks of the *Quercus gracilis–Q. graciliformis* group, retain their dominant role amid the amentiferous taxa. Like in the Pshekha Formation, *Carpinus* spp. is dominant amid Betulaceae of the beds. Single pollen grains correspond to *Rhus* sp., *Acer* sp., *Myrica* sp., *Olea* sp., and some other taxa. Pollen of *Boehlensipollis hohli*, the characteristic taxon of the West European Oligocene, is also encountered in the beds.

The lower Morozkina Balka Subformation. The organic-walled phytoplankton (25 to 35 taxa) represents up to 30% of all palynomorphs distinguished in the basal interval of the subformation. Its abundance decreases down to 7.5% in the middle part of the unit and to 13.5% at the top. In rocks displaying the highest content of C_{org} (up to 10%), dominant taxa are *Batiacasphaera* spp. (up to 50–70%), *B. hirsuta, B. micropapillata*, and *B. baculata* inclusive, and *Hystrichokolpoma* spp., including the thin-walled "suppressed" forms of the genus (up to 10%), which suggest that the basin was of the brackish-water type at the time of transition from the carbonate to clayey–siliceous sedimentation. On the basis of index species detected in all samples studied, the unit is attributed to the *Wetzeliella gochtii* Zone. List of other taxa, whose abundance corresponds to 1–5 specimens per a preparation, is as follows: *Achomosphaera* sp., (Samples 25p and 29); *Apteodinium* spp., including *A. emslandensis* (Samples 25p and 29); *Areosphaeridium pectiniforme* (Sample 27); problematic *Bitectatodinium* sp. (Sample 25p); *Cordosphaeridium* spp., including *C. exilimurum* (samples 28p and 29); *Cerebrocysta* sp. (Sample 28p); problematic *Chiropteridium* sp. (Sample 28p); *Cleistosphaeridium ? insolitum* (Sample 25p); *Cribroperidinium* cf. *guiseppei* (Sample 25p); *Dapsilidinium* sp. (Sample 28p); *Distatodinium ellipticum* (Sample 28p); *Deflandrea spinulosa* (Sample 25p); *Deflandrea* sp. (Sample 29); *Chytroeisphaeridia* sp. (Samples 28p and 29); cf. *Glaphyrocysta inculta* (Samples 25p and 28p); *Horologinella* sp. (Samples 25p, 27, and 29); *Hystrichosphaeridium* sp. (Sample 27; up to 7%); *Homotryblium* sp. (Sample 27); *Impagidinium* sp. (Samples 27, 28p, and 29); *Kallosphaeridium* sp. (Samples 25p and 27); *Lejeunecysta hyalina* (Samples 25p and 28p); *Membranophoridium* sp. (Samples 25p and 28p); *Operculodinium* sp. (Samples 25p and 28p); *Paralecaniella indentata* (Sample 27p); *Pentadinium* spp., including *P. laticinctum* (Sample 28p); *Prolixosphaeridium* sp. (Sample 28p); *Rhombodinium draco* (Sample 25p); *Phelodinium* sp. (Sample 25p); *Spiniferites* spp. (Samples 25p, 27, and 28p); *Tectatodinium* sp. (Samples 25p and 29); and *Wetzeliella* sp. (Samples 25p, 27, and 28p).

Sample 126 from the top portion of the lower subformation revealed the abnormally high abundance of *Chiropteridium* spp., including *Ch. partispinatum* and *Ch. mespilanum* (up to 50% of the phytoplankton assemblage), and of *Wetzeliella* spp. (more than 15%), *W. symmetrica* and *W. gochtii* included. This feature is typical of the upper subformation, but cysts of *Batiacasphaera sphaerica, B. baculata*, and thin-walled *Hystrichokolpoma* spp. coexisting with the above forms suggest interlinks of the unit with the underlying Ostracoda Beds. Other cysts occurring as rare specimens represent *Rhombodinium draco, Cordosphaeridium* sp. (*C. exilimurum* included), *Operculodinium* sp.

Selenopemphix spp. (*S. nephroides* included), *Distatodinium* sp., *Impagidinium* sp., *Dapsilidinium colligerum, Pentadinium* sp., *Chytroeisphaeridia* sp., *Hystrichosphaeropsis* sp., *Spiniferites* sp., *Apteodinium* sp., *Heteraulacacysta* sp., *Paucilobimorpha triradiata*, and *Crassosphaera* sp., all characteristic of the subformation in question.

Changes in the phytoplankton assemblages from the lower Morozkina Balka Subformation were under control of hydrological environments in the basin, especially, of salinity fluctuations. In the course of freshening, the assemblage became enriched in components characteristic of the Ostracoda Beds, such as thin-walled *Hystrichokolpoma* spp., *Batiacasphaera* spp., *Horologinella* sp., *Gerdiocysta* sp., and others. The salinity restoration caused their disappearance and the parallel increase of the taxonomic diversity.

In the lower subformation, proportions between principal components of palynospectra are identical to those in the Ostracoda Beds. In the flowering plant spectrum, the abundance of Fagaceae pollen is about 30%; Betulaceae, mainly *Carpinus* sp. and *Alnus* sp., are less abundant. *Engelhardtia* sp. is the dominant taxon of Fagaceae. Pollen of evergreen oaks and associated *Castanopsis* sp., *Corylopsis* sp., *Hamamelis* sp., *Rhus* sp., *Nyssa* sp., Moraceae, *Ilex* sp., and Araliaceae characterizes episodes of warming similar to that of the Ostracoda Beds time. Spectrum of conifers changes with respect to dominant taxa in the middle and upper parts of the subformation, where the Taxodiaceae/Pinaceae ratio is 7–9 : 1. Pollen of Betulaceae, mainly of *Carpinus* sp., turns out to be dominant amid the amentiferous taxa, and *Engelhardtia* sp. together with *Juglans* sp. become preponderant amid Juglandaceae. The coexisting pollen of thermophilic flowering plants is still diverse. Sample 28p yielded abundant specimens of Hamamelidaceae (up to 10%), *Rhus* sp., and Myricaceae (up to 8%). Single specimens of *Nyssa* sp., *Ilex* sp., *Cornus* sp., *Sterculia* sp., Araliaceae, and *Magnolia* sp. are also recorded. Moderately thermophilic flowering taxa are represented by Ulmaceae (4%), *Platanus* sp., and Tiliaceae. Spore specimens are rare.

The upper Morozkina Balka Subformation. Phytoplankton assemblages of the unit include 30 to 50 taxa. *Hystrichkolpoma* and *Batiacasphaera* species abundant in the underlying subformation occur as rare specimens at this level, and the thin-walled forms are identified in the basal interval only. Acritarchs and green algae, especially, *Leiosphaeridia* spp., display a high abundance. The content higher than below is also characteristic of *Apteodinium emslandensis, Lejeunecysta hyalina, L. paratenella*, and *Homotryblium tenuispinosum. Lejeunecysta* spp. are typical of the Chattian deposits (Zaporozhets, 1998), and their combination with *Rhombodinium draco* and *Chiropteridium* spp. suggests that the upper subformation as a whole can be attributed to the upper Oligocene despite the conclusion by Krhovsky *et al.* (1995), who correlated the lower part of the unit with NP 23 Zone and attributed the middle one to NP 24 Zone. The upper Morozkina Balka Subformation corresponds to the first occurrence level of *Lingulodinium machaerophorum* and *Turbiosphaera* sp. Single dinocyst specimens represent *Selenopemphix nephroides, Dapsilidinium simplex, Spiniferites pseudofurcatus, S. ramosus, Horolonginella* sp., *Tectatodinium* sp., *Cordosphaeridium* sp., and *Pentadinium* sp. Acritarchs and green algae of the unit belong to taxa *Tetraporina* sp., *Micrhystridium* sp., problematic *Variachium* sp., and *Crassosphaera* sp.

In the upper part of the subformation, changes in the microphytoplankton composition are insignificant. Its principal dominant components are still *Wetzeliella* spp. (*W. gochtii, W. symmetrica*, and *W. articulata* included), *Deflandrea* spp. (large *D. spinulosa* cysts, *D. andromensis, D. heterophlycta*, and *D. phosphoritica* f. *lata* included), *Chiropteridium* spp., *Cleistosphaeridium* sp., *Paleocystodinium golzowense*, and *Lingulodinium* sp. In sum, they represent at least 70–75% of the phytoplankton assemblage. *Rhombodinium draco* is also present. Single specimens correspond to more than 20 taxa: *Apteodinium* spp. (*A. emslandense* and *A. maculatum* included), *Batiacasphaera hirsuta, Cribroperidinium* sp., *Cometodinium* sp., *Achomosphaera* sp., *Cordosphaeridium* sp., *Cerebrocysta* sp., *Hystrichokolpoma* spp., *Heteraulacacysta* spp. (several forms), *Lingulodinium* sp., *Lejeunecysta globosa, L. hyalina, Dapsilidinium* sp., *Horologinella* sp., *Hystrichosphaeropsis* sp., *Phelodinium* sp., *Selenopemphix nephroides, Operculodinium microtrianum, Pentadinium laticinctum* sp., *Spiniferites* spp., *Thalassiphora pelagica, Phthanoperidinium comatum*, and others. *Ptanospermella* sp. and *Cymatiosphaera* sp. are leading taxa of green algae. As *Rhombodinium draco* and three *Wetzeliella* species still occur in this part of the section, it is attributed to the lower Chattian. At the same time, the increased abundance of *Lingulodinium machaerophorum, Selenopemphix nephroides*, and *Cleistosphaeridium* sp. is remarkable.

Pollen of conifers, especially, of Taxodiaceae, is dominant in the upper subformation (more than 70%). In the lower part of the unit, Betulaceae prevail in the angiosperm group, whereas in the upper one, where pollen of flowering plants is more abundant, the leading taxa amid Fagaceae, Ulmaceae, Anacardiaceae, and thermophilic broad-leaved plants are *Carpinus*, sclerophyllous oaks, and *Rhus*. Thermophilic taxa are *Nyssa* sp., *Hamamelis* sp., *Corylopsis* sp., *Liquidambar* sp., Moraceae, *Ilex* sp., *Aralia* sp., and *Castanopsis* sp. *Boehlensipollis hohli* characteristic of the lower Oligocene in Western Europe still occurs at this level.

The main part of the upper Morozkina Balka Subformation is of the Chattian age, but its basal beds apparently correspond to the Rupelian. It should be noted, however, that *Chiropteridium* forms become

abundant in upper beds of the lower subformation (Sample 126).

The Batalpashi Formation. Phytoplankton of the formation includes up to 40 taxa, though being less diverse in the upper interval. In the lower half of the unit, it is similar in composition to the assemblage from the upper Morozkina Balka Subformation, although *Deflandrea spinulosa* is somewhat more abundant here, and *Phthanoperidinium* forms are more diverse. This part of the sequence is still of the early Chattian age and corresponds, together with the lower unit, to the *Chiropteridium partispinatum–Rhombodinium draco* Subzone of the dinocyst scale. Phytoplankton from the upper part of the formation shows changes in composition. *Rhombodinium draco, Wetzeliella,* and some other form disappear at this level. Taxa dominant here are *Deflandrea spinulosa, Chytroeisphaeridia* sp., and *Selenopemphix nephroides.* Single specimens represent *Cyclopsiella vieta, Heteraulacacysta campanula, Gerlachodinium aechmophorum, Hystrichokolpoma salacia, Homotryblium tenuispinosum, Pentadinium laticinctum, Apteodinium emslandense, Membranophoridium aspinatum, Melitasphaeridium asterium, Spiniferites* spp., *Cleistosphaeridium* sp., and *Operculodinium* sp. This interval of the sequence corresponds to the upper *Chiropteridium partispinatum–Deflandrea spinulosa* Subzone of the late Oligocene.

In the lower and middle part of the unit, only minor variations in composition of spores and pollen are detected. In general, pollen of conifers, *Rhus,* and Betulaceae is somewhat more abundant here than before, and the spore spectrum is more diverse. This may be a consequence of the onset of a cooling event.

The Septarium Formation. Phytoplankton of the formation consists of about 30 taxa. The majority of them are inherited from the underlying Batalpashi Formation, though with certain changes in abundance parameters. *Deflandrea spinulosa* is dominant in the interval spanning the lower two thirds of the formation sequence. Like many specialists, I consider this taxon in a broad sense, i.e., as a group incorporating both the larger tick-walled forms similar to the type species and the smaller spheroid forms with thinner walls. The latter are characteristic of the Oligocene–Miocene boundary beds, and their disappearance is actually coincident with the extinction time of the genus. Exactly these thin-walled forms are typical of the Septarium Formation. It also yielded *Chiropteridium* spp., *Homotryblium tenuispinosum, Operculodinium* spp., *Spiniferites* spp., *Thalassiphora* sp., *Apteodinium* sp., *Cleistosphaeridium* sp., *Cordosphaeridium* sp., and occasional small *Hystrichokolpoma* spp.

In the upper third of the unit interval, *Deflandrea spinulosa* is missing, but *Labyrinthodinium truncatum* characteristic of the Miocene in Europe and *Bitectatodinium* forms appear at this level. Frequently occurring forms are *Selenopemphix* spp., *Operculodinium* spp., and *Cleistosphaeridium* sp. Green algae and acritarchs are represented by *Leiosphaeridia* spp. and *Cymatiosphaera* sp., the taxa indicative of a partial freshening, or of anoxic event in the basin. The greater part of the formation section is of the Oligocene age, but its upper part displaying the appearance of *Labyrinthodinium* forms and lesser abundance of *Deflandrea spinulosa* is already of the early Miocene age. This conclusion fits well into the inference by Krhovsky *et al.* (1995), who attributed, on the basis of nannoplankton distribution, the lower part of the Septarium Formation to the late Oligocene NP 25 Zone and considered the upper one as corresponding to the boundary interval between this and NN 1 Zone.

Spores and pollen from the Septarium Formation are represented by 70–80 taxa. Pollen of conifers is dominant as before, though the abundance of preponderant Taxodiaceae approaches that of Pinaceae. Fagaceae, Betulaceae, Myricaceae, and Ulmaceae are dominant amid taxa of flowering plants. Pollen of *Engelhardtia* sp., *Comptonia* sp., *Quercus* spp. (apparently deciduous forms), *Betula* spp., and *Rhus* sp. is notably concentrated close to the top of the formation. In this instance, it is interesting to note that the high content of the *Engelhardtia* pollen is characteristic of the lower Miocene deposits in the Crimea–Caucasus region and nearby areas of Western Europe (Planderova, 1990).

Despite some problems concerning the exact position of the Oligocene–Miocene boundary in the Belaya River region, it clearly is between the top of the sandy member and the base of the Karadzhalga Formation, likely closer to the latter. Consequently, the greatest part of the Septarium Formation is of the Oligocene age.

The Karadzhalga Formation. Rare species *Deflandrea spinulosa* are still preserved in the phytoplankton assemblage of the formation. As the geological history of the genus *Deflandrea* ends in the Aquitanian (*Proposal...*, 1994), the presence of its species indicate that the host deposits are the pre-Burdigalian in age.

Like in the top portion of the Septarium Formation, phytoplankton of the Karadzhalga level includes abundant *Leiosphaeridia* spp. in association with frequent specimens of *Spiniferites* spp. and less abundant *Lingulodinium* sp., *Systematophora* sp., *Hystrichokolpoma salacium, Pentadinium* sp., *Apteodinium* sp., *Cleistosphaeridium* sp., *Dapsilidinium* sp., *Homotryblium oceanicum, Lejeunecysta* cf. *globosa, Thalassiphora pelagica, Selenopemphix nephroides, Achomosphaera* sp., *Distatodinium* sp., *Fromea* sp., and *Labyrinthodinium truncatum.*

In the spore-pollen assemblage, pollen of Taxodiaceae prevails a little over that of Pinaceae. Fagaceae and Betulaceae (birches, oaks, and hornbeams) are represented by a wide spectrum of species. Juglandaceae are also diverse and include specimens of *Juglans, Carya, Platycarya,* and *Engelhardtia.* Pollen of Ulmaceae, *Comptonia,* and Hamamelidaceae is rather

abundant as well. *Corylopsis* sp., *Fothergilla* sp., *Ilex* sp., and Moraceae are remarkable taxa of thermophilic plants.

The Ol'ginka Formation. This unit is impoverished in phytoplankton, spores, and pollen. As it is characteristic of the lower Miocene deposits, *Spiniferites* spp. is the dominant taxon of the phytoplankton assemblage also incorporating here the less abundant *Apteodinium maculatum, Selenopemphix nephroides, Lejeunecysta paratenella,* and *L. globosa.* Other taxa encountered in the unit are *Dapsilidinium simplex, Lingulodinium* sp., *Operculodinium* sp., *Pyxidiella scabrosa, Hystrichokolpoma rigaudiae, Samlandia* sp., and *Hystrichosphaeropsis* sp. Acritarchs are represented by *Paucilobimorpha* sp. The redeposited phytoplankton specimens correspond to *Chiropteridium* sp., *Deflandrea* sp., and *Thalassiphora* sp.

The list of taxa characterizing the spore-pollen assemblage is as follows: *Pinus* spp., *Picea* sp., Taxodiaceae, *Sciadopitys* sp., *Podocarpus* sp., *Alnus* sp., *Betula* sp., *Carpinus* sp., *Corylus* sp., *Quercus* spp. (pollen of *Quercus graciliformis* type included), *Engelhardtia* sp., *Carya* sp., *Platanus* sp., Ericaceae, Chenopodiaceae, *Ulmus* sp., *Nyssa* sp., *Rhus* sp., and Moraceae.

The Tarkhanian Regiostage. The Tarkhanian clay beds directly resting on the erosion surface of the Maikop Group bear many redeposited phytoplankton forms (*Wetzeliella gochtii, Charlesdowniea clathrata, Chiropteridium* spp., *Deflandrea granulosa, Thalassiphora* spp., *Samlandia chlamydophora,* and others). The bulk of forms occurs, however, *in situ.* These forms are very diverse (over 60 taxa), and dominant amid them are *Spiniferites* spp., *Operculodinium* spp. (up to 50% of the assemblage), and *Impagidinium* sp. Particular species of these genera are *S. ramosus, S. bentorii, S.* cf. *mirabilis, O. israelianum, O. centrocarpum* (mass abundance), *O. tiare,* and *Operculodinium* sp. I (Manum et Boulter, 1989). Coexisting forms are represented by *Batiacasphaera hirsuta, B. sphaerica, Achomosphaera* sp., *Melitasphaeridium asterium, Dapsilidinium* sp., *Concentricysta* sp., *Tectatodinium psilatum, T. pellitum, Lingulodinium machaerophorum, Tuberculodinium vancampoae, Sumatrodinium* sp., *Hystrichosphaeropsis obscura, Riculacysta* sp., *Cribroperidinium* sp., *Distatodinium paradoxum, Invertocysta* sp., *Systematophora placacantha* (and other species of the genus), *Cordosphaeridium inodes, Lejeunecysta oliva, L. sabrina, L. paratenella, Trinovantedinium* sp., *Caligodinium pychnum, Caligodinium* sp., *Rhombodinium* sp., *Apteodinium piasecki, Pentadinium goniferum, Selenopemphix nephroides, Homotryblium tenuispinosum, Hystrichokolpoma* sp., and cf. *Oligosphaeridium* sp.

Taxodiaceae are dominant in the spore-pollen assemblage of the regiostage, and Juglandaceae (*Carya* sp., *Juglans* sp., *Engelhardtia* sp., *Pterocarya* sp., and *Platycarya* sp.) are the main taxa of the angiosperm group. Single pollen grains represent *Comptonia* sp., *Aesculus* sp., *Ulmus* sp., Ericaceae, Moraceae, Hamamelidaceae, *Fothergilla* sp. (cf. *F. gracile*), *Ilex* sp., *Rhus* sp., and *Tilia* sp. Coexisting spores belong to *Gleichenia* sp. and *Lygodium* sp. The characteristic Miocene species *Momytes punctatus* Nagy is also present. In addition, the spectrum includes herbaceous Chenopodiaceae and fruticose *Comptonia* forms, and, like the phytoplankton assemblage, it includes the redeposited pollen. The latter points once again to the erosion of upper horizons of the Maikop Group prior to accumulation of the Tarkhanian deposits.

REGIONAL DINOCYST ZONATION

The suggested regional zonation of organic-walled phytoplankton (Fig. 2) is based on the presented data for the Paleogene–lower Miocene section of the Belaya River basin. The distribution of most important taxa is shown in the table, and some of them are pictured in Plates I and II.

The upper half of the middle Eocene (Bartonian) is divided into the *Areosphaeridium diktyoplokus* and *Rhombodinium perforatum* zones. Characteristic species of the lower zone are *Cerodinium diebelii, Cordosphaeridium exilimurum, Diphyes colligerum, Lejeunecysta globosa, Manumiella* sp., and *Wetzeliella echinulata.* Distinctive features of the upper *Rhombodinium perforatum* Zone are the predominance of *Areosphaeridium arcuatum* over *A. dictyoplokus* and the appearance of additional *Rhombodinium* species (*R. draco* and *R. rhomboideum*), *Microdinium reticulatum, Melitasphaeridium asterium,* and *Heteraulacacysta leptalea.*

The *Charlesdowniea clathrata angulosa* Zone of the upper Eocene, which corresponds to the range of the Belaya Glina Formation, is well correlative with D 12 Zone of the West European dinocyst scale (Costa and Manum, 1988; Köthe, 1990). Characteristic taxa of the unit are *Achomosphaera sagena, Cannosphaeropsis* sp., *Cordosphaeridium funiculatum, Rottnestia borussica, Fibrocysta* sp., *Turbiosphaera* sp., *Wilsonidium* sp., and *Tytthodiscus beloglinensis.* Species *A. diktyoplokus, Rottnestia borussica, Tytthodiscus beloglinensis,* and some other taxa do not cross the Eocene–Oligocene boundary. The *Microdinium reticulatum* Beds distinguished at the base of the Belaya Glina Formation in some localities adjacent to the Ciscaucasia, e.g., in the northern Ergeni area (Zaporozhets, 1998), represent a member transitional between the Kuma and Belaya Glina formations of the Belaya River region.

The lower Oligocene is divided into the *Phthanoperidinium amoenum, Wetzeliella symmetrica,* and *Wetzeliella gochtii* zones.

The *Phthanoperidinium amoenum* Zone spans the basal Member I of the Pshekha Formation, which is known under the name *Planorbella* Beds. This zone is fairly well corresponding to D 13 Zone of Western

Plate I

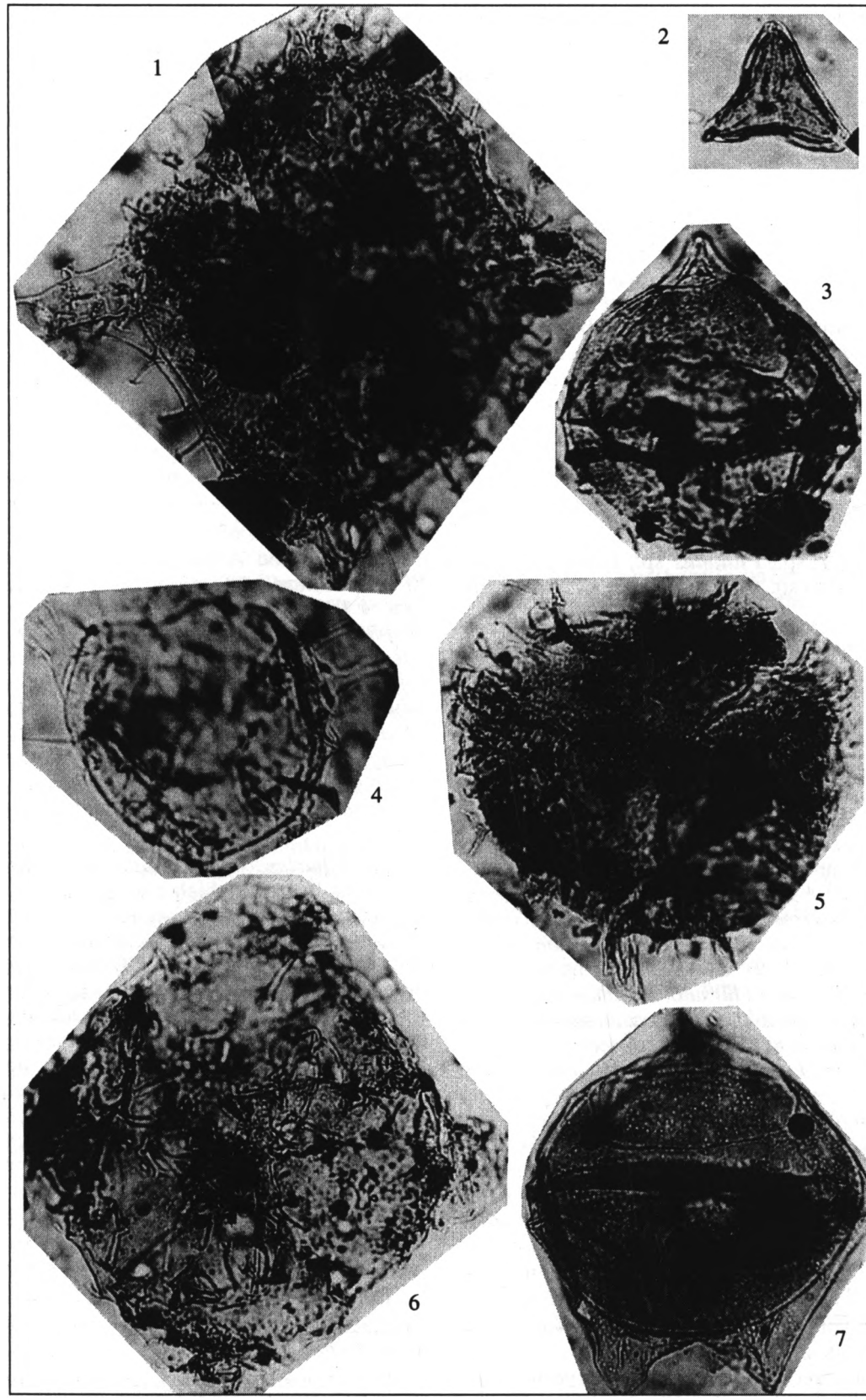

Plate I. Oligocene and early Miocene dinocysts from the Maikop Group (Belaya River region of western Ciscaucasia).
(1) *Wetzeliella symmetrica* Weiler, 1956, Sample 25, Ostracoda Beds; (2) *Boehlensipollis hohli* W. Kr., 1959, Sample 24, Ostracoda Beds; (3) *Deflandrea spinulosa* Alberti, 1959, Sample 19, Septarium Fm.; (4) *Chiropteridium mespilanum* (Maier, 1959) Lentin et Williams, 1973, Sample 126; Morozkina Balka Fm.; (5) *Chiropteridium partispinatum* (Gerlach, 1961) Brosius, 1963, Sample 126, Morozkina Balka Fm.; (6) *Wetzeliella gochtii* Costa et Downie, 1976, Sample 36, Morozkina Balka Fm.; (7) *Deflandrea spinulosa* Alberti, 1959, Sample 19, Septarium Fm.

Europe. *Homotryblium floripes, Deflandrea phosphoritica, Gonyaulacacysta* sp., *Lejeunecysta hyalina*, and *Cribroperidinium tenuitabulatum* represent distinctive taxa of the unit.

The *Wetzeliella symmetrica* Zone includes Member II and a greater portion of Member III of the Pshekha Formation. Its lower boundary marks the first occurrence level of the index species. In the West European scale, an equivalent of the unit is D Subzone 14a spanning the lower half of the Rupelian excluding the basal beds of the stage. The zonal assemblage is represented by *Lejeunecysta* and *Selenopemphix* species (*S. nephroides* and *S. coronata* included) in association with *Phthanoperidinium* forms (*P. comatum, P. amoenum, P. echinatum*), *Areosphaeridium pectiniforme, Adnatosphaeridium multispinosum, Cleistosphaeridium, Dapsilidinium simplex*, and *Systematophora placacantha*. The zone is also marked by the disappearance of *Areosphaeridium arcuatum*.

In the Belaya River sections, the upper part of Member III of the Pshekha Formation together with the lower Morozkina Balka Subformation is distinguished as corresponding to the *Wetzeliella gochtii* Zone, whose equivalent in the West European Rupelian Stage is D Subzone 14b. Nannoplankton zonal units correlative with the *W. gochtii* Zone are the upper part of NP 22 Zone and the greatest part of NP 23 Zone below its terminal beds. The first occurrence level of index species marks the lower boundary of this dinocyst zone. The variable composition of zonal assemblage suggests frequent changes in hydrological environments (salinity, temperature, and oxygen saturation in the basin of that time). When the basin salinity was approaching the normal level, the assemblage mainly consisted of *Deflandrea* species, *D. spinulosa (typicus)* included, and they coexisted with *Cleistosphaeridium (?) insolitum, Cordosphaeridium exilimurum, Cribroperidinium* cf. *guissipei, Distatodinium ellipticum, Glaphyrocysta inculta, Membranophoridium* sp., and some other forms.

Within the range of *W. gochtii* Zone, the Ostracoda Beds interval can be considered as an individual intrazonal *Batiacasphaera–Hystrichokolpoma* Beds, which yielded a wide spectrum of species representing both genera (in particular, *B. hirsuta, B. baculata*, and *B. micropapillata* as well as *Hystrichokolpoma* spp. including seven thin-walled morphotypes apparently exemplifying endemic species of a short stratigraphic range). These basic taxa coexist with *Gerdiocysta* sp., *Paralecaniella indentata, Horologinella* sp., *Spiniferites elongatus, S. mirabilis*, cf. *Leberidocysta* sp., *Stoverocysta* forms, and representatives of the *Glaphyrocysta–Cyclonephelium–Adnatosphaeridium* group, which are especially characteristic of the Ostracoda Beds in the Kuban sections. The beds in question are undistinguished in sedimentary sequences characterizing the North Atlantic and Tethyan basins of normal salinity.

The *Chiropteridium partispinatum* Zone of the upper Oligocene corresponds in range to D 15 Zone of the West European dinocyst zonation and is divisible into two subzones. The *Chiropteridium partispinatum–Rhombodinium draco* Subzone spans the interval of the upper Morozkina Balka Subformation and the greater part of the Batalpashi Formation. It corresponds to the terminal level of NP 23 Zone plus NP 24 Zone. The *Ch. partispinatum–Deflandrea spinulosa (non typicus)* Subzone incorporates upper beds of the Batalpashi Formation together with lower, middle, and (partially) upper members of the Septarium Formation. This subzone closely corresponds in range to NP 25 Zone.

In addition to index species and *Rh. draco*, the lower subzone showed presence of *Lejeunecysta* and *Deflandrea* species associated with *Apteodinium emslandense, A. maculatum, Operculodinium microtrianum, Pentadinium laticinctum, Thalassiphora pelagica, Phthanoperidinium comatum*, and *Selenopemphix nephroides*; it also marks the first occurrence level of *Lingulodinium machaerophorum*. The distribution range of three *Wetzeliella* species (*gochtii, symmetrica*, and *articulata*) extends up to the boundary with the overlying subzone. Accordingly, it is necessary to ascertain the boundary position between the lower and upper Oligocene in the Belaya River sections, where *Ch. partispinatum*, the index species of the latter substage, initially appears at the Ostracoda Beds level, and *W. gochtii* occurs even in basal beds of the Batalpashi Formation. At present, it is very likely that the aforementioned boundary runs inside the lower 10-m-thick bed sequence of Member I of the upper Morozkina Balka Subformation.

Indicative taxa of the upper subzone are *Deflandrea spinulosa* morphotypes represented by small thin-walled forms. They associate with *Homotryblium tenuispinosum, Selenopemphix nephroides, Chytroeisphaeridia* sp., *Chiroeipteridium partispinatum, Cyclopsiella vieta, Heteraulacacysta campanula, Gerlachodinium aechmophorum, Operculodinium* sp., *Spiniferites* sp., and some other forms.

The *Labyrinthodinium truncatum* Zone of the lower Miocene spans terminal beds of the Septarium Formation and the whole interval of the Karadzhalga Formation. The unit is correlated with D 16 Zone of the West

Plate II

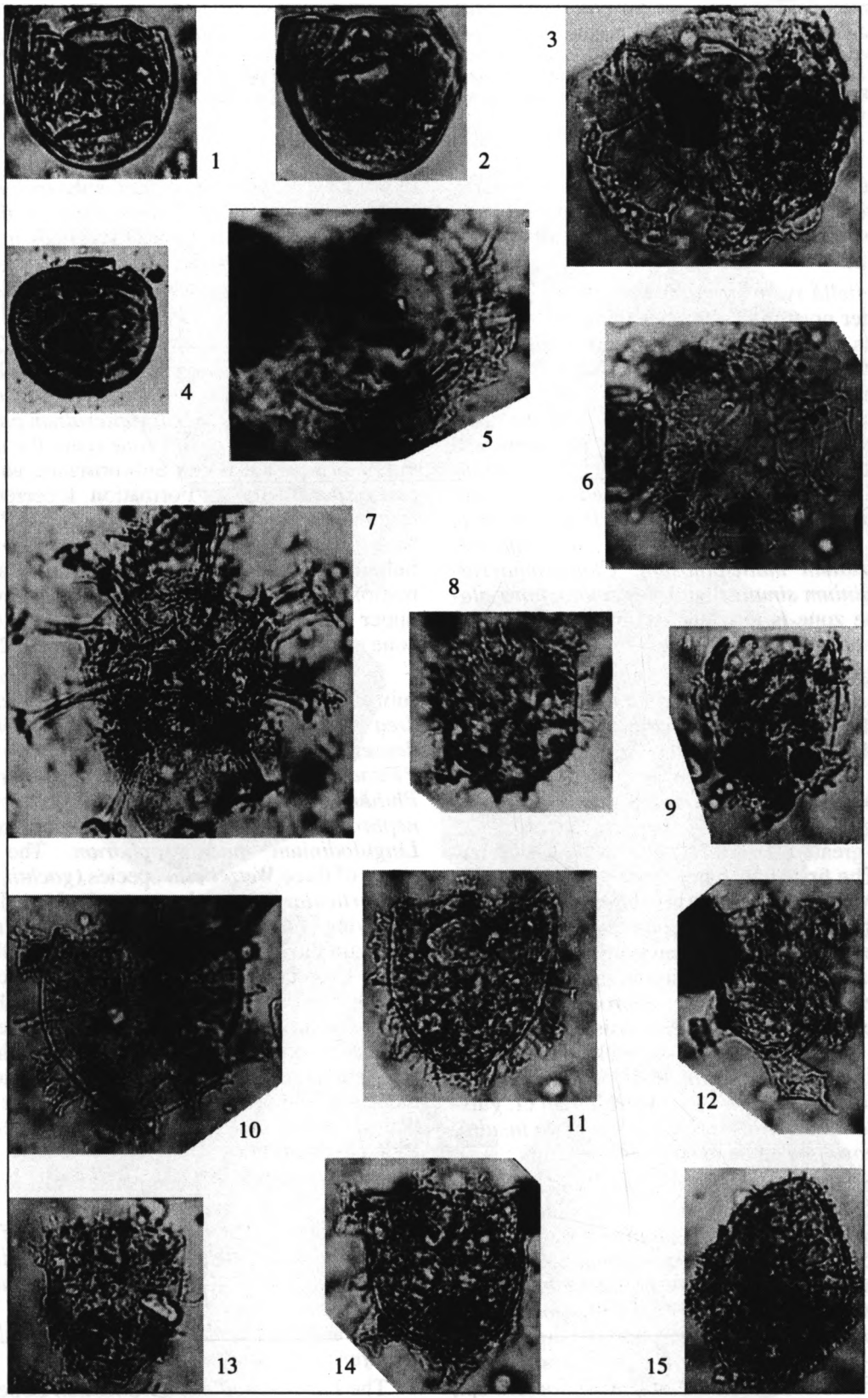

Plate II. Oligocene and early Miocene dinocysts from the Maikop Group (Belaya River region of western Ciscaucasia).

(1, 2) *Batiacasphaera sphaerica* Stover, 1977, Sample 25, Ostracoda Beds; (3) *Glaphyrocysta exuberans* (Deflandre et Cookson) Stover et Evitt, 1978, Sample 25, Ostracoda Beds; (4) *Batiacasphaera micropapillata* Stover, 1977, Sample 36, Morozkina Balka Fm.; (5, 6) *Labyrinthodinium truncatum* Piasecki, 1980, samples 33 from Septarium Fm. (5) and 568 from Karadzhalga Fm. (6); (7) *Homotryblium tenuispinosum* Deavey et Williams, 1966, Sample 15, Morozkina Balka Fm.; (8–14) *Hystrichkolpoma* spp., diverse thin-walled morphotypes, Sample 25, Ostracoda Beds; (15) *Operculodinium* sp., Sample 25, Ostracoda Beds.

All images are shot under magnification ×600; preparations are stored at the Laboratory of Paleobotany, Geological Institute RAS.

European dinocyst zonation, although its index species is recorded only in the lower Miocene deposits of Denmark. The *Deflandrea spinulosa* abundance is notably decreased in this zone, where *Bitectatodinium* forms appear for the first time and *Selenopemphix, Operculodinium*, and *Cleistosphaeridium* species are characteristic of its microphytoplankton assemblage. The zone is correlative to NN 1 Zone. Krhovsky *et al.* (1995) consider the lower part of the zone, where *D. spinulosa* is still a component of fossil remains, as an interval transitional between NP 25 and NN 1 Zones.

The overlying Tarkhanian deposits yielded the very diverse assemblage of *Spiniferites, Operculodinium*, and *Impagidinium* species, some middle Miocene representatives of these genera included. Other taxa characteristic of the first half of Miocene and occurring at this level are *Tuberculodinium vancampoae, Lingulodinium machaerophorum*, and *Hystrichosphaeropsis obscura*. In the Mediterranean region, similar assemblages characterize the boundary interval between the lower and middle Miocene, but most typical species of the middle substage are missing here.

CONCLUSION

(1) Data on the studied organic-walled phytoplankton from the Paleogene–lower Miocene sections of the Belaya River region and known distribution patterns of these microfossils in other areas of Ciscaucasia (Zaporozhets, 1993) are used to elaborate the regional dinocyst zonation suggested in this paper (Fig. 2). This zonation is better substantiated as compared to the regional scale formerly elaborated by Andreeva-Grigorovich (1991) for the Crimea–Caucasus region. Compositions of zonal assemblages are characterized by the more complete lists of indicative taxa, and ages of local stratigraphic units are better defined. The distinguished dinocyst zones are correlated with nannoplankton zonal units, whose characteristics were simultaneously studied by Krhovsky in the same collection of samples.

(2) The abundance of palynomorphs of higher plants is found to be higher in deposits accumulated during the periods of hydrosulfuric contamination of sedimentary basins (marls of the Kuma Formation, the lower Morozkina Balka Subformation, the Batalpashi and Karadzhalga formations).

(3) The decreased abundance of coniferous pollen in the Belaya Glina Formation points to a warm subtropical climate of the late Eocene and to the low contrast of relief forms in land areas adjacent to the basins. The commencement of cooling in the second half of the late Eocene is marked by the grown abundance of Pinaceae pollen in the *Propeamussium fallax* Beds.

(4) A warmest period of the Oligocene corresponds to the second half of the Rupelian, when the diverse pollen of thermophilic flowering plans became abundant.

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Early Pleistocene Faunas of Small Mammals in Eastern Europe

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Abstract—Materials from many lower Pleistocene localities of small mammal remains elucidate principal stages of the fauna evolution in eastern Europe and enable the biostratigraphic subdivision of host deposits on the basis of evolutionary changes in small mammal communities and their coordination with geological and geochronological data. Several assemblages and evolutionary phases of early Pleistocene mammals are distinguished. The first occurrence levels of new taxa (mostly of genera and species of Arvicolidae) are used to detect boundaries between assemblages. The established geochronological subdivisions of the Pleistocene are coordinated with those of West European scales. The initial period of the early Pleistocene and an early phase of the Odessa fauna of small mammals correspond to the first occurrence level of arhizodont vole *Allophaiomys deucalion*. Later faunas of the Odessa assemblage are characterized by the appearance of arhizodont vole genera *Prolagurus (P. ternopolitanus)* and *Lagurodon (L. arankae)*, who had teeth lacking cementum. These taxa coexisted with *Allophaiomys deucalion*. The more progressive Nogaisk faunas of small mammals (Topachevskii, 1965) included the evolved species *Allophaiomys pliocaenicus*, which coexisted with *Prolagurus ternopolitanus* and *Lagurodon arankae*. The next stage of evolution was marked by origin of first small lemmings *Prolagurus pannonicus*. The term Kairy faunas is suggested to be appropriate for assemblages of this stratigraphic level (derived from the Zapadnye Kairy stratotype locality in the Dnieper River lower courses). The Kairy faunas also include voles *Allophaiomys pliocaenicus, Mimomys savini*, and *Mimomys pusillus*. These faunas are correlative in age with the paleomagnetic Jaramillo Subchron. More progressive assemblages with first *Microtus (Terricola)* and *Microtus (Stenocranius ?) hintoni* have been already described as the Morozovka faunas (Aleksandrova, 1976). Advanced *Mimomys* forms, *Allophaiomys pliocaenicus nutiensis*, and *Prolagurus pannonicus* are characteristic of this stratigraphic level. Successive Petropavlovka faunas of small mammals correspond in time to the end of the early Pleistocene and to the Matuyama Chron. Their index forms are first voles *Microtus* ex gr. *oeconomus*. The advanced *Mimomys* species (*M. savini, M. pusillus*) of the Petropavlovka faunas coexisted with dominant *Prolagurus pannonicus, Eolagurus argyropuloi*, and *Microtus (Stenocranius) hintoni*. The early Tiraspol faunas correlative with the uppermost level of the Matuyama Chronozone are distinct because of newcomers *Microtus arvalinus* and *Prolagurus posterius*.

Key words: *early Pleistocene, small mammals, faunal assemblages.*

INTRODUCTION

A great volume of information about fossil small mammals of the early Pleistocene from Eastern Europe has been accumulated during the last decades. Bone remains of this age are widespread in the central and southern parts of the Russian plain (Fig. 1). Faunas of the late Pliocene–early Pleistocene have been investigated by Shevchenko (1965), Topachevskii (1965, 1973), Sukhov (1970), Aleksandrova (1976), Topachevskii and Skorik (1977), Topachevskii *et al.* (1987), Markova (1982, 1990a, 1990b, 1992), Agadzhanyan and Glushankova (1986), Kazantseva (1990), Agadzganyan and Markova (1993), Rekovets (1994), and Tesakov (1995).

Data available at present characterize more than sixty lower Pleistocene localities of small mammals in the territory of Eastern Europe. This data base was used to elucidate the sequential evolutionary changes in communities of small mammals and to discriminate their principal assemblages and phases of evolution during the early Pleistocene.

The analysis was performed with due account for data of geology, malacology, palynology, and geochronology. Morphological and evolutionary changes in phylogenetic lineages of Rodentia, Lagomorpha, Insectivora, and primarily in the family Arvicolidae were investigated in detail and used to substantiate stratigraphic boundaries between the assemblages of small mammal and phases of their evolution.

Faunas, which are direct ancestors of the oldest early Pleistocene assemblages, have been described by Gromov (1939, 1948) from the Psekups locality. As it was detected later, some assemblages of small mammals from this locality were lacking arhizodont voles and incorporated *Archidiskodon meridionalis meridionalis, Mimomys* ex gr. *reidi-pusillus*, progressive *Mimomys pliocaenicus, Borsodia fejervaryi*, and *B. arankoides* (Gromov, 1948; Aleksandrova, 1977; Vangengeim *et al.*, 1990; Tesakov, 1995; see Fig. 1).

Paleomagnetic scale	Stratigraphic scale		Ma	Mammals					
				Assemblage of large mammals	Assemblage of micro-mammals	Phase of micro-mammal assemblage	*Taxa of the first occurrence level*	East European sites	West European sites
B	Pleistocene	middle	0.78	Tiraspolian	Tiraspolian	Early Tiraspolian	*Microtus arvalinus, Prolagurus posterius*	Semibalka-1, Litvin Shamin	Zuurland (27–37 m), Stranska Skala,
		lower	1.00	Tamanian	Tamanian	Petropavlovka	*Microtus (Microtus) ex. gr. oeconomus*	Karai-Dubina, Petropavlovka, Log Krasnyi, Novo-Troitskoe, Semibalka-3, Port-Katon,	Bavel (?)
						Morozovka	*Microtus (Stenocranius ?) hintoni* *M. (Terricola) sp.*	Morozovka 1 (= Cherevichnoe 1), Luzanovka, Bolshevik 2, (Bed 4)	Meiningen
J						Kairy	*Prolagurus pannonicus*	West Kairy, Ushkalka, Korotoyak (Ostrogozhsk Fm.)	Monte Peglia (t. r.), Monte Peglia (b. r.), Bagur 2, Deutch Altenburg 30A, Ostramosh 14
						Nogaisk	*Allophaiomys pliocaenicus*	Roksolany, Nogaisk, Tarkhankut, Korotoyak 3 (upper bed), Chishmikioi, Akkulaevo (Demsk and Davlekanovo horizons)	Pirro Nord, Mac Rambo, Betfia VII/1 (upper bed), Deutch Altenburg 2C1,
				Psekups	Odessa	late Odessa	*Lagurodon arankae, Prolagurus ternopolitanus*	Kryzhanovka (upper bed), Log Denisov, Korotoyak 3 (lower bed), Khadzhimus, Chortkov, Uspenka, Melekino, Zhevakhova Gora (5, 9)	Betfia IX, Betfia X, Zabia Cave, Zuurland (42–46 m), Mokrá 1, Betfia XIII
O		Pliocene	2.00			early Odessa	*Allophaiomys deucalion*	Tiligul, Mikhailovka 1	Zuurland (50–58 m), Kolinany 3, Kamyk, Kadzielnya, Villany 5
					Psekups	Saratov	*Borsodia ex. gr. fejervaryi*	Saratov (Psekups), Svapa, Liventsovka (upper bed)	Zuurland (91–95 m) Tegelen

Fig. 1. Stratigraphic position of sites with bones of small mammals: (B) Brunhes Chronozone; (J) Jaramillo Subchronozone; (O) Olduvai Subchronozone.

According to Vangengeim *et al.*, (1990), bone remains of small mammals are confined in the Psekups locality to an upper part of the middle Akchagylian (Kuyalnik) beds, namely, they occur inside the Matuyama Chronozone of reversed polarity. These remains were termed as the Saratov fauna (Aleksandrova, 1977). Pevzner (1989) located the upper Kuyalnik boundary between the Olduvai and Gills subchrons and established that the Kuyalnik local stage spans the entire Akchagylian Stage and basal beds of the Apsheronian. On the basis of molluscan fauna, the Akchagylian assemblage of small mammals was found to be older than the Olduvai Subchron, i.e., older than 1.9 Ma. In the opinion of the aforementioned authors, there is a chance that the remains of arhizodont voles may be present in the Psekups locality.

Micromammal faunas of a similar evolutionary level were also discovered from the upper unit of the Liventsovka section near Rostov-on-Don (Aleksandrova, 1976) and in the Svapa locality of the Dnieper upper courses (Agadzhanyan and Kalutskaya, 1976). In Western Europe, faunas of small mammals from the Tegelen and Zuurland localities (in the last case, level of 91–96 m) of the Netherlands (Kolfschoten and Meulen, 1986; Kolfschoten, 1988, 1990) presumably correspond to the same stratigraphic level.

ODESSA ASSEMBLAGE OF SMALL MAMMALS

The Odessa assemblage of small mammals was initially discovered by Shevchenko at the Kryzhanovka locality. The assemblage includes the diverse *Mimomys* forms and is characterized by the appearance of the first arhizodont voles (*Allophaiomys* and *Lagurodon*) either bearing, or lacking cementum in their teeth (Shevchenko, 1965). This assemblage is correlative with faunas of large mammals incorporating *Archidiskodon meridionalis meridionalis* and also may be correlated with the late phase in evolution of the Psekups assemblage of large mammals. In addition, the Odessa assemblage is coeval to the lower–middle Guri deposits of the Black Sea region and to the middle Apsheronian beds of the Caspian Sea region (Nikiforova and Aleksandrova, 1991). Stages in the evolution of the Odessa fauna can be considered in detail on the basis of data obtained recently.

Early Phase of the Odessa Fauna

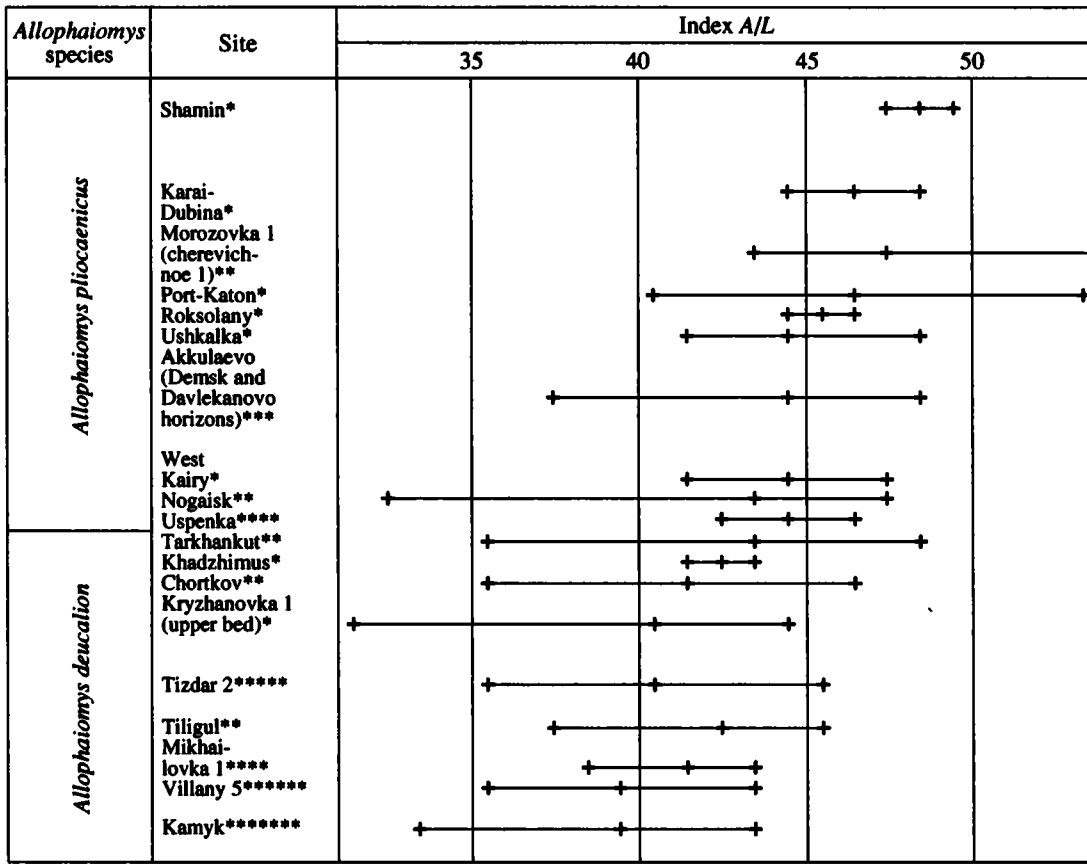
The oldest micromammal faunas with remains of arhizodont voles *Allophaiomys* (*A. deucalion*) are described from the locality Mikhailovka 1 in the Dnieper River upper courses (Agadzhanyan and Glushankova, 1986) and from the Tiligul site near the Tiligulskii Bay of the northern Black Sea (Topachevskii and Skorik, 1977; Rekovets, 1994; see Fig. 1). Arhizodont voles from genera *Prolagurus* and *Lagurodon*, which are lacking cementum in their teeth, are missing from this faunas. *Borsodia* voles of that time demonstrate a high level of hypsodonty (Tesakov, 1993). Supergenus *Mimomys* is represented in this case by the same species, which are characteristic of the Psekups faunas. Lower anterior molars of *Allophaiomys* voles are characterized by a simple paraconid structure and by low A/L index (less than 42, Fig. 2) typical of *A. deucalion* (Agusti, 1991). The enamel structure of *Allophaiomys* forms is archaic, because it is wider in posterior areas of lower dental prisms. Widening (bulging) of enamel is characteristic of the alveolar side of some teeth. Similar faunas were discovered at the localities Villany V in Hungary and Kamyk in Poland (Kretzoi, 1956; Kowalski and Nadachowski, 1990).

The stratigraphic position of these faunas is disputable. Until the recent time, the first occurrence level of *Allophaiomys* forms was dated as approximately corresponding to 1.6 Ma, and many sites of corresponding bone remains were correlated with the lower–middle Guri and lower–middle Apsheronian deposits of the Black and Caspian seas, respectively.

The *Allophaiomys* taxa have not been detected yet among micromammal remains from the Psekups stratotype locality, and I consider it reasonable to use the term "Odessa fauna" for assemblages of small mammals incorporating the archaic *Allophaiomys* forms. It is also noteworthy that the *Archidiskodon meridionalis meridionalis* assemblages of large mammals existed for a longer period of time, and their stratigraphic range spans the distribution intervals of both the Psekups and Odessa faunas of small mammals. The early Odessa phase seems to correspond to the Eburonian of Western Europe. Faunas of this age are known from the level of 50–58 m of the Zuurland locality of the Netherlands (Kolfschoten, 1988, 1990) and occur in several other Eburonian sites, e.g., in Kolinany 3 and Včeláre 5, 10, and 10B of Czechia (Fejfar and Horaček, 1983, 1990), in Kamyk and Kadzielnya of Poland (Garapich and Nadachowski, 1996), in Neuleiningen 15 of Germany (Koenigswald and Tobien, 1990), and in Villany 5 of Hungary (Kretzoi, 1956; Janossy, 1986).

Late Phase of the Odessa Fauna

The late phase of the Odessa micromammal fauna is characterized by the first appearance of arhizodont voles from genera *Prolagurus (P. ternopolitanus)* and *Lagurodon (L. arankae)*. In the Russian plain, faunas of this evolutionary level occur in a series of localities, such as the following: the upper bed of the Kryzhanovka site (the stratotype of the Odessa micromammal assemblage distinguished by Shevchenko); Zhevakhova Gora site (beds 5 and 9) near Odessa; Uspenka, Korotoyak 3 (lower bed), and Log Denisov sites in the Don River basin; Khadzhimus site at the Dniester River; Chortkov site in the Seret River basin; and Melekino site of the Sea of Azov coast (Shevchenko, 1965; Markova, 1990a; Agadzhanyan, 1992; Michailescu and Markova, 1992; Agadjanian and Kazantseva, 1994;



* This work;
** (Rekovets, 1994);
*** (Sukhov, 1970);
**** (Agadzhanyan, personal communication);
***** (Tesakov, personal communication);
****** (Kretzoi, 1969);
******* (Garapich and Nadachowski, 1996).

Fig. 2. A/L index values for *Allophaiomys* M_1 from the early Pleistocene localities in Eastern Europe.

Rekovets, 1994). Teeth of *Allophaiomys* forms from these sites show an archaic structure typical of *A. deucalion*. Liguridae are represented by oldest species *Prolagurus ternopolitanus* and *Lagurodon arankae*. Geographic ranges of lagurines were not identical. The distribution area of *Prolagurus ternopolitanus* was wider, and bone remains of this taxon occur up to 52°N, whereas the localities with *Lagurodon arankae* remains are situated in the southernmost areas of the Russian plain, i.e., in the northern Black Sea and Sea of Azov coasts (Figs. 3a, 3b). This implies a certain difference in the ecological adaptability of these animals. Lemmings *Prolagurus ternopolitanus* were distinctly more adaptive, and *L. arankae* forms inhabited only the arid and hot steppe landscapes. In contrast to the sites of Romania, where the first occurrence levels of *Lagurodon arankae* and *Prolagurus ternopolitanus* are asynchronous (Terzea, 1995), both taxa appeared almost concurrently and coexisted in time within the Russian plain despite the evident difference in their geographic ranges.

Remains of *Lemmus kowalskii* from the Chortkov site are unique in the micromammal faunas under consideration. So far, this is the only find from the lower Pleistocene localities of Eastern Europe.

Mimomys ex gr. *reidi-pusillus* and last *M. pliocaenicus* are present in the faunas amid representatives of this genus. The Khadzhimus section also yielded remains of *Mimomys pitymyoides*, the very rare species of Eastern Europe, where it is recorded in this and Psekups sites only. According to results of thermoluminiscent dating, alluvium of the Dniester terrace IX bearing the micromammal fauna is 1300 ± 350 Ma old (Kulikov and Chepalyga, 1985). Faunas of similar evolutionary level are also known from the following localities: Betfia XII site of Romania bearing *Lagurodon arankae* and early *Allophaiomys* forms (Terzea, 1984, 1995); Mocrá site of Slovakia (Fejfar and Horaček,

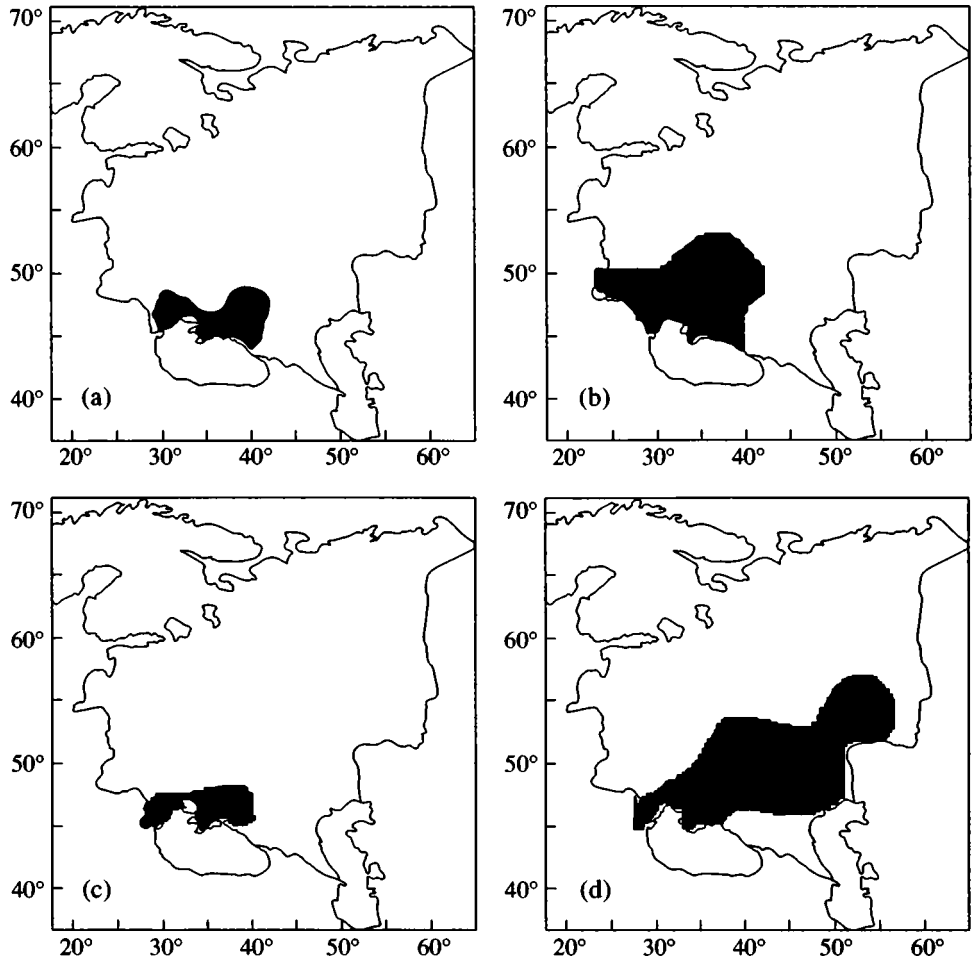


Fig. 3. Distribution areas of lemmings in Eastern Europe: (a) *Lagurodon arankae* and (b) *Prolagurus ternopolitanus* during the late Odessa phase; (c) *Lagurodon arankae* and (d) *Prolagurus ternopolitanus* at the Nogaisk fauna time.

1983); Zabia Cave site of Poland (Garapich and Nadachowski, 1996); level of 42–46 m in the Zuurland site of Netherlands (Kolfschoten, 1988); Nagyhärsányhegy 2 of Hungary (Kretzoi, 1961; Janossy, 1986); and Venta Micena of Spain (Agusti, 1991). They all are attributed to the early Biharian (Kretzoi, 1961).

TAMAN FAUNA

Archidiskodon meridionalis meridionalis is the index form of the Taman fauna. Large mammals characteristic of this fauna existed for a rather long period of time corresponding to the accumulation interval of the upper Guri deposits in the Black Sea region correlative with the upper Apsheronian beds of the Caspian Sea area (Nikiforova and Aleksandrova, 1991). Calibration with reference to the paleomagnetic scale shows that the time range of the fauna is approximately equal to 300 ka and lasted from 1.1 to 0.8 Ma (Vangengeim *et al.*, 1991). The existence period of the *Archidiskodon meridionalis meridionalis* is divisible into four distinct intervals termed as the Nogaisk, Kairy, Morozovka, and Petropavlovka evolutionary phases of small mammals.

Nogaisk Phase

The Nogaisk fauna of micromammals and synonymous phase of their evolution were originally distinguished at the Nogaisk locality of the northern coast of the Sea of Azov (Topachevskii, 1965). In the Russian plain (Fig. 1), faunas of this type also occur in the Tarkhankut site (Topachevskii, 1973), in the upper bed of the Korotoyak 3 locality (Agadzhanyan, 1992), in the Chishmikioi burial (Shushpanov, 1977), and in the Demsk and Davlekanovo horizons of the Akkulaevo site (Sukhov, 1970).

The Nogaisk assemblage includes progressive *Allophaiomys* voles (*A. pliocaenicus*). Index A/L for M_1 of the *Allophaiomys* forms is over 42 (Fig. 2). Remains of lemmings *Lagurodon arankae* and *Prolagurus ternopolitanus* (= *"P. praepannonicus"*) together with bones of *Allophaiomys* forms are dominant in all localities (Topachevskii, 1965, p. 129, Fig. 35). Rekovets (1994)

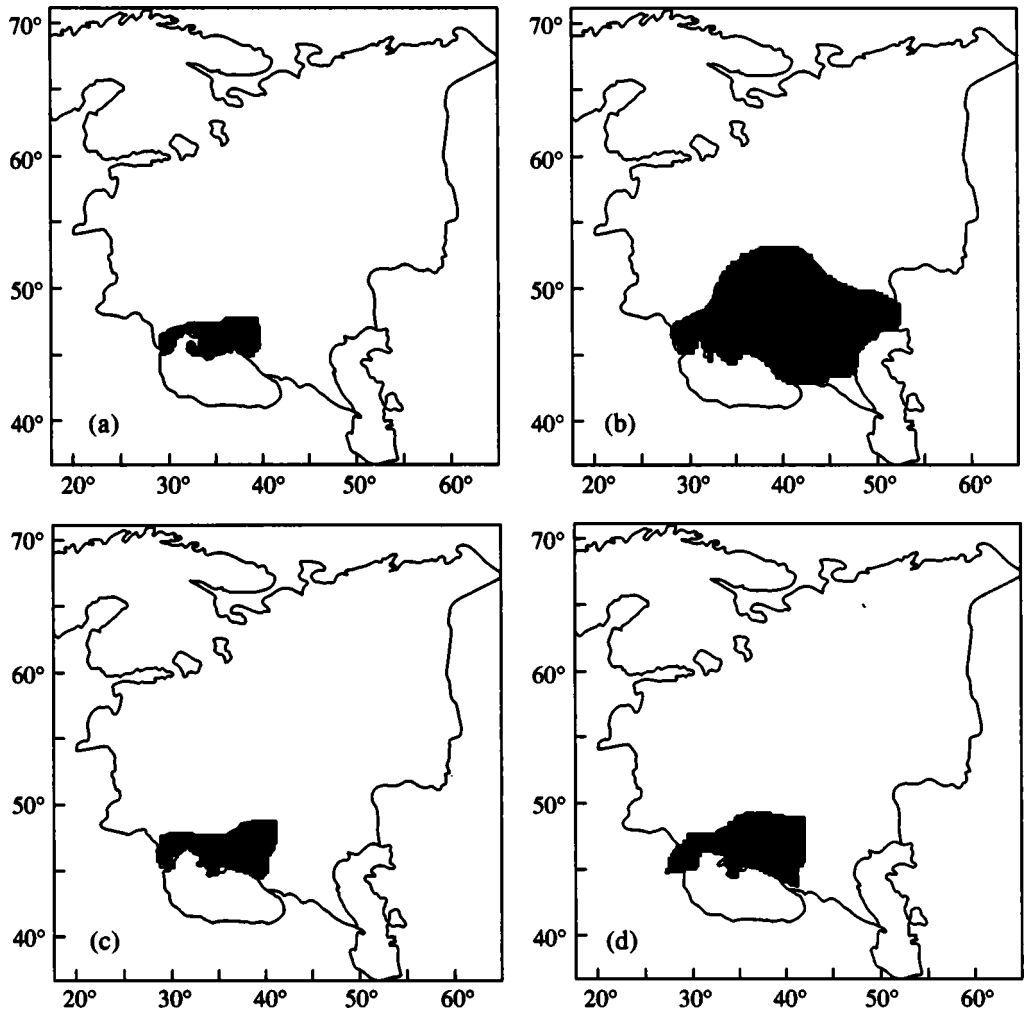


Fig. 4. Distribution areas of lemmings in Eastern Europe: (a) *Lagurodon arankae* and (b) *Prolagurus ternopolitanus* at the Kairy fauna time; (c) *Lagurodon arankae* and (d) *Prolagurus ternopolitanus* at the Morozovka fauna time.

revised the lagurine remains of the *Prolagurus* lineage, presented a new description of these micromammals, and classed them with *P. pannonicus*. In my opinion, however, the peculiar morphology of *Prolagurus* teeth from the Nogaisk and, especially, from the Tarkhankut sites that was pictured by Topachevskii in his books of 1965 (p. 129, Fig. 35) and of 1973 (p. 169, Fig. 53), is incompatible with that of the type dental series of *P. pannonicus* from the Püspökfürdo site (Kormos, 1930), because dental crowns in the paraconid complex M_1 are well differentiated in many teeth of the series. Indices M_1 estimated by Rekovets in his book also suggest that *Prolagurus* teeth from Nogaisk and Tarkhankut localities differ significantly from lagurine teeth of younger taxa of this phylogenetic lineage (e.g., species from the Kairy site and others). At the same time, it is clear that lagurines from the Nogaisk and Tarkhankut localities can be attributed to most progressive representatives of *P. ternopolitanus*.

During the Nogaisk phase, geographic ranges of *L. arankae* and *P. ternopolitanus* were again different. *L. arankae* inhabited the extreme south of the Russian plain in contrast to a wider distribution area of *P. ternopolitanus* (Figs. 3b, 3c). *Borsodia* remains are scarce in the Nogaisk faunas. *Mimomys* ex gr. *reidi-pusillus* occurs in the assemblages of micromammals, where remains of vole species *M. savini* (= *"intermedius"*) are abundant. Peculiar forms of genus *Ellobius (E. paleotalpinus, E. tauricus, E. tarkhancutensis)* and species *Clethrionomys sokolovi* are also characteristic of the stratigraphic level in question (Topachevskii, 1965; Rekovets, 1994). Archaic yellow lemmings *Eolagurus argyropuloi* represent micromammals that first appeared at this level as a result of prochoresis (Topachevskii, 1973). West European faunas from the site Včeláre 3B/1, where they incorporate *Allophaiomys pliocaenicus, Prolagurus "praepannonicus," Lagurodon arankae*, and *Mimomys reidi-pusillus* (Horaček, 1985), are similar to the Nogaisk faunas.

Sites Pirro Nord and Pietraffita in Italy (Masini and Torre, 1990) may also represent this phase of micromammal evolution.

Kairy Phase

The next stage in evolution of small mammals is marked by appearance of lagurines *Prolagurus pannonicus* displaying a progressive structure of dental apparatus and representing the direct descendants of *Prolagurus ternopolitanus*. Sites with these faunas numerous in the Russian plain (Fig. 1) have also been discovered in the Dnieper (West Kairy and Ushkalka localities), Dniester (Roksolany locality), and Don (Korotoyak locality and Ostrogozhsk Formation) river basins (Markova, 1982; Kazantseva, 1990; Michailescu and Markova, 1992; Iosifova and Krasnenkov, 1994; Krasnenkov, 1994). In addition to *Prolagurus pannonicus*, *Allophaiomys pliocaenicus* demonstrating a progressive morphotype of teeth (enamel of the *Microtus* type, index A/L = 44–45) is another characteristic form of these faunas (Fig. 2). These forms coexisted with *Lagurodon arankae, Clethrionomys sokolovi*, and *Eolagurus argyropuloi*. Genus *Mimomys* is represented mainly by *M. savini* and *M. pusillus*. Distribution areas of *Prolagurus pannonicus* and *Lagurodon arankae* were different like at the time of the Nogaisk faunas. During the Kairy phase, the geographic range of *P. pannonicus* was wider than that of the last species and extended through southern and central areas of the Russian plain (Figs. 4a, 4b).

I consider it reasonable to discriminate this fauna from other early Pleistocene assemblages of small mammals and to term this evolutionary phase after the Zapadnye Kairy reference site, where old alluvial deposits are exposed in the section of the Kairy terrace in the Dnieper River lower courses (Lebedeva, 1972). In this section, the alluvial deposits bearing bones of small mammals are overlain by the 20-m-thick sequence of loess–soil deposits. The section was described by Markova (1982a, pp. 19–22). Pevzner, who studied paleomagnetic characteristics of the sequence, demonstrated that the Brunhes–Matuyama boundary is located here above two fossil soil horizons separated by a loam interlayer and resting on alluvium with fossil bones of micromammals (see in Velichko *et al.*, 1976).

In the Korotoyak section, where paleomagnetic parameters of rocks were studied by Semenov (1994), the bed with micromammal bones inside the Ostrogozhsk Formation shows the normal polarity and corresponds to the Jaramillo Subchron. On the basis of paleomagnetic data, the existence time of Kairy faunas can be dated back to the period of 1.0–1.1 Ma ago. The Kairy faunas are correlative with West European faunas from sites Deutch Altenburg 2C, Betfia IX, and Bagur 2, where *Allophaiomys pliocaenicus, Prolagurus pannonicus*, and *Lagurodon arankae* have been encountered (Rabeder, 1981; Terzia, 1995; Lopez *et al.*, 1976). These sites are attributed to the early Waalian phase.

Morozovka Phase

Specific features of Morozovka faunas marking the next evolutionary level consist in further morphological transformations of *Allophaiomys* voles and in the appearance of progressive voles *Microtus (Stenocranius ?) hintoni* and *M. (Terricola) (?)* sp. Aleksandrova (1976) was first to distinguish faunas of this evolutionary level in the type section Morozovka 1 located at the left-hand side of the Khadzhibeiskii Bay. Later, faunas of this type were discovered at the Port-Katon site (Markova, 1982, 1990b). Rekovets (1994) reported additional finds at the Morozovka 1 (Cherevichnoe 1) site and on similar faunas recovered from the Luzanovka and Bolshevik 2 sections (Fig. 1) situated in Odessa.

Late forms of *Allophaiomys* and *Mimomys* (*M. savini* and *M. pusillus*) are not abundant in the Morozovka faunas. Lemmings are represented in this case by *Prolagurus pannonicus, Lagurodon arankae*, and *Eolagurus argyropuloi*. First and second lagurine forms of this phase again inhabited different geographic areas (Figs. 4c, 4d). Index A/L for M_1 of the *Allophaiomys* species is over 47 (Fig. 2). The Morozovka faunas were coeval with Tamanian assemblages of large mammals incorporating *Archidiskodon meridionalis tamanensis*, whose remains were recovered, in particular, from the Port-Katon section. In western Europe, approximately the same level yielded the Meiningen fauna with remains of *"Pitymys"*-like voles originally described as *Microtus (Allophaiomys) pliocaenicus nutiensis* (Heinrich, 1982) and later classed with *Microtus thenii* (Maul, 1990). In my opinion, these voles are more advanced than late *Allophaiomys* forms. They are at the evolutionary level close to that of early *Terricola* = "Pitymys" species from the Russian plain, for instance, they are very similar in morphology to *Microtus (Terricola)* sp. from the Port-Katon locality (Markova, 1990b, Fig. 15, image 9).

Petropavlovka Phase

The Petropavlovka faunas are most advanced in evolution, as compared to other early Pleistocene micromammals, and include the new species *Microtus (Pallasiinus)* ex gr. *oeconomus*. The faunas are named after the Petropavlovka locality in the Don River basin (Krasnenkov and Agadzhanyan, 1975; Aleksandrova, 1976). Faunas of a similar evolutionary level were discovered near the village of Karai-Dubina in the Dnieper River lower courses (Markova, 1982; Rekovets, 1994) and at the Log Krasnyi site near the village of Uryv, the Don River lower courses (Kazantseva, 1990).

Key forms of the Petropavlovka faunas are lemmings *Prolagurus pannonicus* and *Eolagurus argyropuloi* associated with voles *Microtus (Stenocranius ?)*

hintoni. Other characteristic forms are rhizodont voles *Mimomys savini*, *M. pusillus*, and late *Borsodia*, but they are scarce. Few *Allophaiomys* teeth from above sites are progressive in morphology, and their enamel is differentiated in the *Microtus* mode. Index A/L for M_1 of *Allophaiomys* forms occurring in the Petropavlovka faunas is about 48 (Fig. 2). In Western Europe, this level of micromammal evolution is represented by the Holstein fauna from Czechia (Musil, 1966; Fejfar and Horaček, 1983).

The time period in question is marked by the appearance of an elephant form more progressive than the Tamanian *Mammuthus trogontherii* (Dubrovo, 1985). The elephant bones were discovered from the Mikhailovka terrace IV of the Dniester River, and the Brunhes–Matuyama paleomagnetic reversal was detected in alluvial deposits of the terrace.

According to analysis of paleomagnetic data on the Petropavlovka fauna sites, the bone-bearing beds correspond in this case to the top of the Matuyama Chronozone of the reversed polarity. In the Karai-Dubina section, the paleomagnetic reversal is recorded in the floodplain facies of alluvium resting on the fluvial bed with remains of small mammals (Velichko, *et al.*, 1983). The Petropavlovka faunas are terminal in the succession of the early Pleistocene assemblages of micromammals in the Russian plain.

Faunas more advanced in evolution than the Petropavlovka faunas correspond in age to the topmost portion of the Matuyama Chronozone marking the commencement of the middle Pleistocene. The Shamin site in the Don River lover courses is the principal one bearing faunas of this type. In addition to *Microtus* ex gr. *oeconomus*, these faunas include another species of the genus, namely *M. arvalinus*, and both associate with *Microtus (Stenocranius ?) hintoni* and *Prolagurus posterius*, the descendant of *P. pannonicus* (Fig. 1). Other species of the Shamin faunas are represented by *Mimomys pusillus, Eolagurus* ex gr. *argyropuloi*, and *Allophaiomys pliocaenicus nutiensis*. The faunas were attributed to the group of the early Tiraspol assemblages (Markova, 1982, 1990a).

Faunas from the Litvin site in the Tamanskii Peninsula, where they occur in deposits of the Chauda transgression, are similar to the Shamin faunas in the evolutionary level and taxonomic composition. This site corresponds, however, to the Brunhes Chronozone of normal polarity. The site yielded remains of *Mimomys savini, Allophaiomys pliocaenicus nutiensis, Prolagurus pannonicus, Eolagurus simplicidens gromovi, Microtus (Stenocranuis ?) hintoni*, and *Microtus (Microtus) arvalinus* (Markova, 1990a). Accordingly, the early Tiraspol faunas seem to be synchronous with the boundary interval between the Matuyama and Brunhes chronozones.

CONCLUSION

As it is evident from the above analysis, the early Pleistocene micromammal faunas of Eastern Europe are divisible into several assemblages on the basis of changes in the phylogenetic lineages and appearance of new taxa, first of all voles and some other forms, demonstrating particular phases of their evolution.

(1) The earliest Pleistocene faunas of small mammals are characterized by the appearance of the first arhizodont voles of the genus *Allophaiomys (A. deucalion)*. Key taxa of the faunas are the rhizodont vole genera *Mimomys* and *Borsodia* (*M.* ex gr. *reidi-pusillus, B. fejervaryi*, and *B. arankoides*). The arhizodont voles with teeth lacking cementum are missing from these faunas attributed to the early phase of the Odessa micromammal assemblage. The coeval fauna of large mammals includes *Archidiskodon meridionalis meridionalis*.

(2) The next stage in evolution of small mammals is marked by coexistence of most archaic arhizodont voles *Prolagurus ternopolitanus* and *Lagurodon arankae* with *Allophaiomys deucalion. Archidiskodon meridionalis meridionalis* is still the index species in assemblages of large mammals. These faunas characterize the late Odessa phase in development of micromammals.

(3) The Taman faunal assemblage with *Archidiskodon meridionalis tamanensis* corresponds to several stages in the evolution of small mammals. *Allophaiomys pliocaenicus* is the index taxon of the Nogaisk micromammal faunas correlative with the early Taman phase of terriofauna. As before, the archaic lagurines *Prolagurus ternopolitanus* and *Lagurodon arankae* were members of the Nogaisk faunas.

(4) The first occurrence level of an advanced Laguridae form *Prolagurus pannonicus* marks a new stage in the evolution of small mammals. In addition to this lemming form, characteristic species from this stratigraphic level are represented by *Lagurodon arankae, Allophaiomys pliocaenicus, Mimomys pusillus*, and *M. savini*. Faunas of the level correspond in time to the Jaramillo Subchron. They are suggested to be termed the Kairy faunas after the name (Zapadnye Kairy) of their reference site.

(5) The newly appeared vole taxa *Microtus (Stenocranius ?) hintoni* and *Microtus (Terricola)* sp. are indicative of the next stage in development of the early Pleistocene micromammals. Fossil assemblages with these taxa are named the Morozovka faunas (Aleksandrova, 1976).

(6) Taxonomic composition of the latest Petropavlovka faunas of early Pleistocene micromammals is characterized by appearance of voles *Microtus* ex gr. *oeconomus*. Principal taxa of the faunas are voles *Microtus (Stenocranius ?) hintoni* and lemmings *Prolagurus pannonicus* and *Eolagurus argyropuloi*. The genus *Mimomys* of rhizodont voles is represented by low-abundant bone remains (*M. savini* and *M. pusil-*

lus). Species of the genus *Allophaiomys* are very rare and represented by the progressive *A. pliocaenicus nutiensis*. Sites of these faunas are confined to an upper part of the Matuyama Chronozone. The Petropavlovka faunas are transitional between the Tamanian and Tiraspolian assemblages of micromammals.

(7) The commencement of the middle Pleistocene is characterized by the early Tiraspolian faunas with *Microtus arvalinus* and *Prolagurus posterius*. These faunas are correlative with the termination of the Matuyama Chronozone.

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Planktonic Foraminifers from Sediments of the Southeastern Okhotsk Sea (the Last 20000 years)

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Abstract—Analysis of the taxonomic composition of planktonic foraminifers and oxygen isotope composition in their shells from the sedimentary core section V34-90, taken from the eastern slope of the Akademii Nauk SSSR Rise in the southeastern Okhotsk Sea, were used to investigate climatic and hydrologic changes in the basin for the last 20000 years according to radiocarbon dating. Periods of glaciation (20000–12500 years ago), deglaciation (12500–800 years ago) with a clearly expressed cooling event (Younger Dryas, about 10800 years ago), and postglaciation (last 8000 years) were distinguished.

Key words: *planktonic foraminifers, glaciation, deglaciation, postglaciation, Younger Dryas, isotoopic events, the Sea of Okhotsk.*

INTRODUCTION

Detailed investigations of past climatic changes in high latitude basins require the biostratigraphic scale to be of a higher resolution and various events to be dated precisely. From the viewpoint of detailed biostratigraphy and paleoecology, the Sea of Okhotsk is studied inadequately. Despite certain published results of micropaleontological research of recent and Quaternary sediments of the region (Jouze, 1962; Saidova, 1961; Kurihara, 1968; Lipps, 1966), the sporadic sampling of sections and few age determinations restricted the possibility to subdivide sedimentary sequences in detail, to date and correlate various paleohydrologic events.

The studied sedimentary section V34-90 was taken from a slope of the Akademii Nauk Rise in the south of the Sea of Okhotsk central area. Relatively high sedimentation rates and frequent sampling provided grounds to reconstruct (at the precision level of 1000 years) changes in the composition of planktonic foraminiferal assemblages, their dwelling conditions, and accumulation of carbonate shells on sea floor. The revealed distribution patterns of planktonic foraminifers were correlated with oxygen isotope parameters and radiocarbon dates for *Neogloboquadrina pachyderma* (Ehrenberg) sin.

Surface Water Hydrology

A peculiar feature of the Sea of Okhotsk hydrologic regime is the specific behavior of its water mass (Moroshkin, 1966), which largely determines microplankton development. In May, the spring warming brings about the division of the upper layer into the surface and deeper water masses. The latter is represented by a cold intermediate layer within the depth interval between 30–40 and 150–200 m, where it has negative temperatures and is spread all over the sea in spring, summer, and autumn. Temperatures in the meridional section of the basin along 150°E close to the Akademii Nauk Rise area are as follows: in May, the temperature in the near-surface layer (0–200 m) is 0–1°C; in August, the layer warms up to 10–12°C down to the depth of 30 m, whereas the temperature in the layer ranging from 30–40 to 200 m remains low (0–1°C). In the autumn–winter season, as the surface layer cools down and convection intensifies, the surface water mass disappears and waters in the 0–200 m interval again become isothermal.

Materials and Techniques

Core section V34-90 was taken during Leg 34 of R/V "Vulkanolog" from the southeastern slope of the Akademii Nauk Rise at a depth of 1590 m (coordinates: 48°49.9'N, 150°27.6'E; Fig. 1). The section was 333 cm long.

The total abundance of planktonic foraminifers and the amount of individual species were calculated in the rock fraction more than 0.05 mm. The results are recalculated per 1 g of dry sediment. In total, 35 samples have been analyzed (table).

Planktonic foraminifers were encountered in all studied samples; their concentrations range from several tens to 7000 specimens per 1 g of dry sediment. The total abundance of planktonic foraminifers, the abundance of *N. pachyderma* and *Globigerina bulloides* d'Orbigny, and the relative content of individual

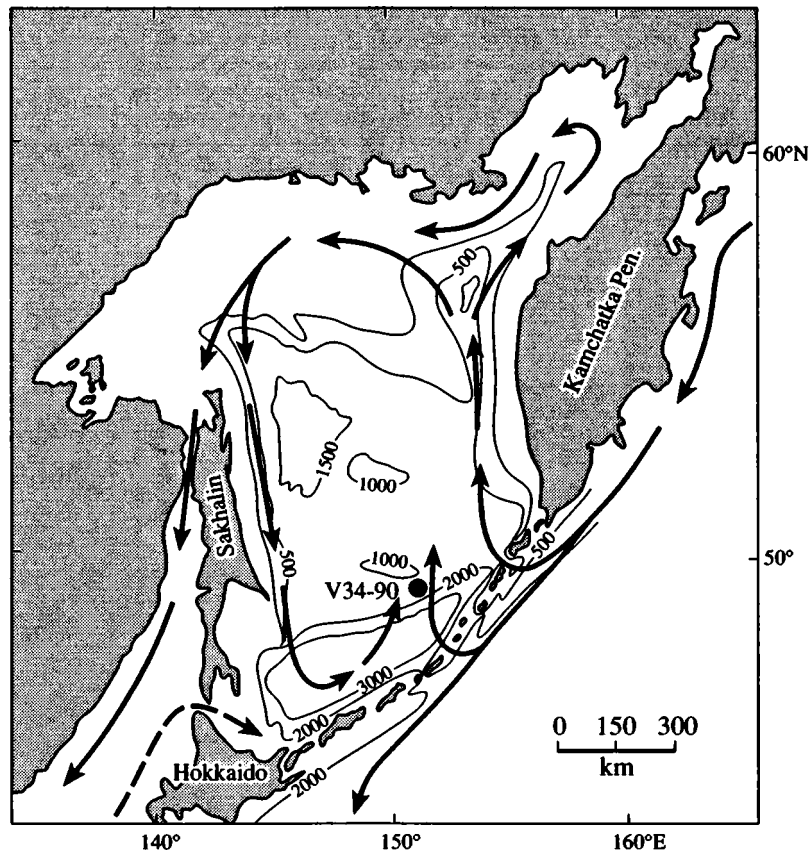

Fig. 1. Position of core site V34-90 and the present-day circulation of surface water in the Sea of Okhotsk (solid arrows indicate cold currents, dotted—warm).

species in the section of Site V34-90 are given in the table and in Fig. 2.

The oxygen isotope parameters in $CaCO_3$ of *N. pachyderma* shells, 0.25–0.1 mm in size, were measured at the Pacific Institute of Oceanology, Far East Division, Russian Academy of Sciences. Age determinations were performed at the Livermore National Laboratory (California, USA) by J. Southon using the accelerator mass spectrometry technique (Gorbarenko *et al.*, 1998). The radocarbon datings on calcium carbonate in *N. pachyderma* shells, marked by an asterisk in the table, are reduced for the reservoir-corrected factor by subtracting 1100 years (Southon *et al.*, 1993).

Lithology

Core section V34-90 was taken in the south of central area of the sea, where bottom sediments are characterized by the highest silica content. In the granulometric aspect, the section is represented by pelitic silts uniform in composition and consisting of 85% to 100% of silt grains less than 0.01 mm in size. The content of silty fraction (0.1–0.01 mm) decreases from 10–15% in the lower part of the section to about 5% at the 160-cm level, and to less than 1% in the upper 110 cm. In clastic terrigenous deposits of the lower half of the section (depth interval of 333–140 cm), the content of biogenic components ($CaCO_3$ and SiO_2) varies from 7 to 20%. Higher (140–115 cm), there are transitional sediments, whose terrigenous material is gradually substituted upward by biogenic substance, the content of which ranges from 24 to 43% (from 20 to 35% SiO_2 and from 6 to 9% $CaCO_3$). The upper 110-cm interval is composed of diatom ooze containing 40–55% of SiO_2 and 3–8% $CaCO_3$. The $CaCO_3$ content varies throughout the section from less than 1% to about 9%, and the highest values are recorded in the following intervals: 220–210 cm (5–6%), 190–185 cm (7.2–9.2%), 140–115 cm (up to 9%), and at a depth of 40 cm (8%). The C_{org} concentration varies through the section, from 0.4 to 8% in its lower part (333–220 cm) and from 0.5 to 1.5% in its upper part (Fig. 3).

Analysis of the Distribution of Planktonic Foraminifers

Planktonic foraminifers in the section are mostly represented by polar species *N. pachyderma* sin.; subpolar *G. bulloides* also occurs in great amounts. The ratio between these two species; the content of *Globigerinita quinqueloba* (Natland); the total abundance of planktonic foraminifers; and the distribution of

Sampling intervals, isotopic age of sediments, and major quantitative parameters of planktonic foraminifers distribution in core section V34-90

Depth, cm	Age, years	Abundance, specimens/g	*N. pachyderma*	*N. pach.*	*G. bulloides*	*G. quinqeloba*	*G. uvula*	*G. scitula*	*G. glutinata*
			sin., %	dex, %	%	%	%	%	%
0–3		46	94		2	2	2		
12		300	95		4	1			
20–23		130	92		5	3			
30–33	1000	650	88		8	4	+		
40–43		140	91		4	5			
50–53		430	78	1	10	10		1	
60–63	2170*	190	91		7	1	1		
70–73		340	87		5	5	3		
80–82		110	87	3	8	1	1	1	
90–93	3080*	1150	79		10	8	1		1
100–103		870	71		14	11	3		
110–113		360	79		15	4			
120–122	4000	3100	97		2	1	1	1	
130–132		4700	88	+	8	2	1		
140–142	6000	2480	93		4	1	+	1	1
150–152	7000?	1370	86	1	10	1	1		
160–162	8000	1570	73	1	24	1	+	+	+
170–172		2390	78	2	19	1			
174	8770*								
180–182		4600	65	+	34	1	+	+	
184	9230*								
194		1000	80		19	1	+		+
200–203		270	86	1	9	1		3	
204		40	94		5	1			
210–213		5440	70	+	28	2	+	+	
220–223	12180*	6940	62	2	30	5	+	1	
230–233		1860	82	5	10	3			
240–243	14000	1440	89	1	7	3			
250–253	15000	1190	85		11	3			
260–263	15950*	1280	79		17	3			
270–273		900	71	7	18	4			
280–283		900	66	1	31	2			
290–293		1550	80	1	16	2			
300–303	19830*	1600	80	2	15	3			
310–313		380	80	+	20	1			
320–323		430	87		11	1			
330–333		380	93		5	1			

Note: Symbol "+" indicate presence of shells.

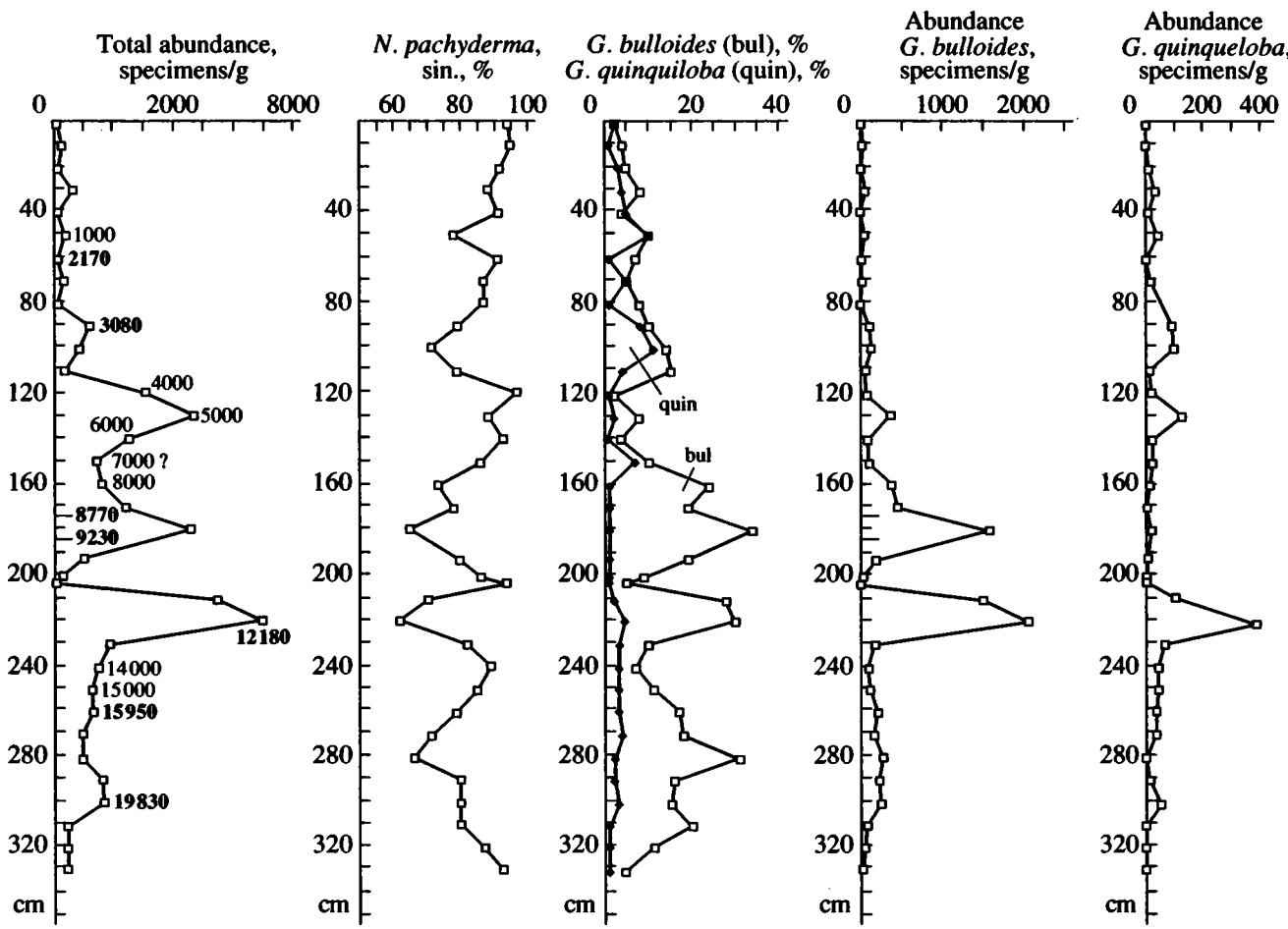

Fig. 2. Changes in the total abundance of planktonic foraminifers, *G. bulloides* and *G. quinqueloba* shells, and in percentages of *N. pachyderma, G. bulloides*, and *G. quinqueloba* throughout the core section V34-90. The indicated age values (years) are based on radiocarbon dating of *N. pachyderma* sin. shells (bold underlined) or interpolated (normal font).

Globorotalia scitula (Brady), *Globigerinita glutinata* (Egger), and *Globigerinita uvula* (Ehrenberg) are used as major characteristics for paleoceanological interpretations.

Changes in the oxygen isotope composition of *N. pachyderma* shells and radiocarbon dating of the shells allowed us to distinguish glacial (333–224 cm), deglacial (224–160 cm), and postglacial (160–0 cm) sediments in the core section V34-90.

In the glacial terrigenous sediments of the section lower part (333–224 cm), polar *N. pachyderma* sin. species makes up 71–93%; *G. bulloides* shells are also abundant (7–20%); *G. quinqueloba* (1–2%) and *N. pachyderma* dex. (less than 1 to 7%) occur in subordinate quantity.

In sediments of this interval, the content of subpolar species *G. bulloides* and of relatively thermophilic dextral *N. pachyderma* greatly varies. In particular, in the intervals of 283–280 and 273–270 cm, the content of *G. bulloides* amounts to 31 and 18%, respectively. Thermophilic dextral *N. pachyderma* is also present in the assemblage; its content here is highest in the section (7%). Increasing proportion of these species is supposed to reflect a minor warming episode about 17000 years ago (Peteet *et al.*, 1992). In the interval of 253–240 cm, the content of *G. bulloides* becomes 2–3 times lower, and *N. pachyderma* dex. may even disappear at this level (perhaps in response to the glacial maximum recorded 15000–14000 years ago in the highest $\delta^{18}O$ values (Fig. 3)).

Based on the absolute concentration of planktonic foraminifers and $CaCO_3$ content, the section of glacial deposits is divided into two parts. The sediments of the lower part (333–310 cm) with less than 1% of $CaCO_3$ and low absolute concentrations of foraminiferal shells (380–430 specimens/g) displaying dissolution marks were formed more than 20 000 years ago. In the upper part of the glacial deposits (303–223 cm; age span of 20000–13000 years), the absolute abundance of planktonic foraminiferal shells increases to 900–1860 specimens/g and $CaCO_3$ content amounts to 1–2%. The abundance of shell fragments in glacial deposits is low and suggests that the water corrosive influence on cal-

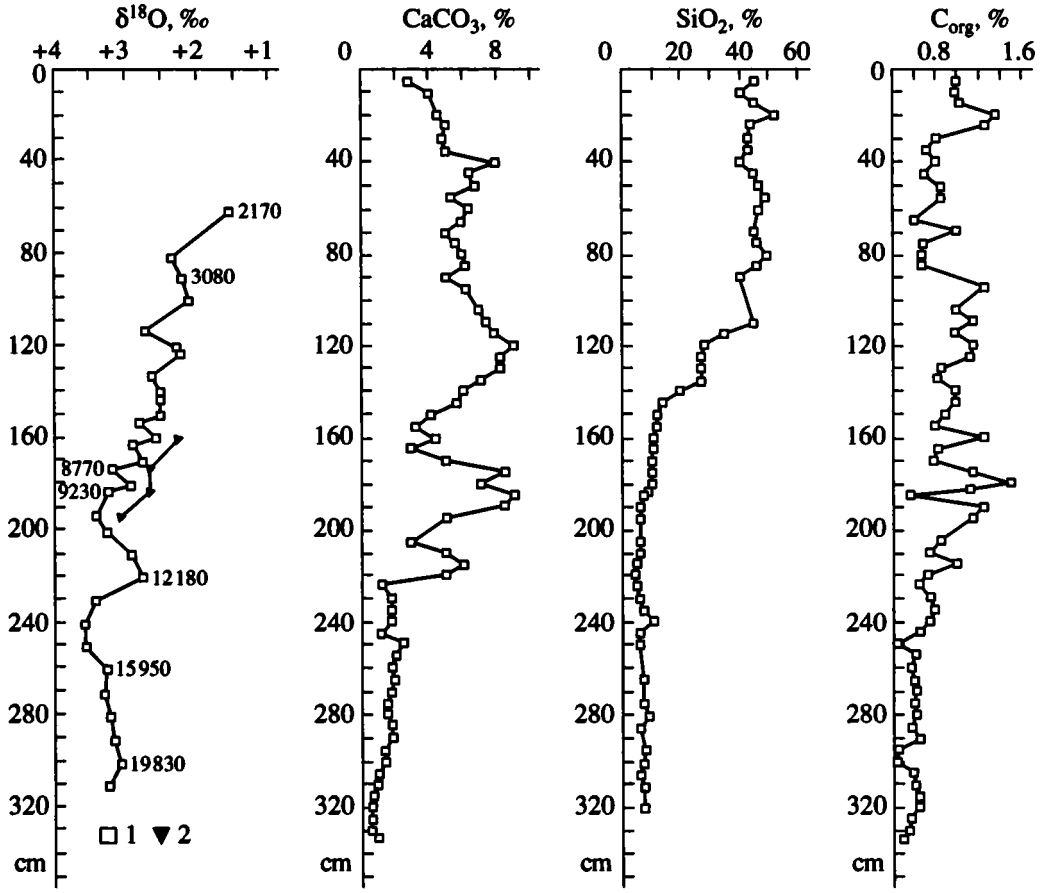

Fig. 3. Variation curves of ^{18}O in *N. pachyderma* sin. (1) and *G. bulloides* (2) shells, and of the calcium carbonate ($CaCO_3$), amorphous silica (SiO_2), and organic carbon (C_{org}) content (wt% for dry sediment) in core section V34-90 (indicated radiocarbon dates for *N. pachyderma* sin. in the left column are after Gorbarenko *et al.*, 1998).

cium carbonate was rather weak. In general, the low content of $CaCO_3$, SiO_2, and C_{org} in glacial deposits (less than 2, 10, and 0.8%, respectively) indicate a low productivity of the surface water layer and an intensive terrigenous influx at that time. An average sedimentation rate was 10.3–10.6 m/k.y. (Fig. 4).

Terrigenous sediments of the middle part of the section (223–160 cm), which correspind to the deglacial epoch (12500–8000 years ago), are characterized by the increased diversity of the planktonic foraminiferal assemblage up to 7 species due to the appearance in small quantities of *G. scitula, G. glutinata*, and *G. uvula*. The role of subpolar species *G. bulloides* grows in this interval, and the abundance of their shells amounts to 20–30%. The abundance of foraminiferal shells reaches its maximum (about 6940 specimens per 1 g of sediment) with a simultaneous general increase of carbonate content in sediments.

Three intervals are identified in this part of the section according to changes in foraminiferal assemblages and isotopic characteristics of shells. The interval of 223–210 cm, which is characterized by a high absolute abundance of foraminiferal shells and the dominance of the *N. pachyderma* sin., *G. bulloides*, and *G. quinqueloba*, coincides with a clearly expressed negative excursion of $\delta^{18}O$ values and with the first phase of ice sheet thawing or Termination 1A (Gorbarenko *et al.*, in press). It is the interval where *G. uvula* and *G. scitula* shells are recorded for the first time. The maximum (for the entire section) abundance of foraminiferal shells (6940 specimens/g) was detected in the layer of 223–220 cm corresponding in age to 12180 years. The $CaCO_3$ content here increases to 5%, whereas that of C_{org} remains as low as in the glaciation sediments. In the layer of 213–210 cm, the amount of foraminiferal shells decreases parallel to the increasing of the C_{org} content to 1% and $CaCO_3$ to 6% due to growing abundance of foraminiferal shells in sediments. This is indicative of the progressing dissolution of shell carbonates, which occurred later than the beginning of deglaciation.

A sharp decline in the abundance of planctonic foraminifers (up to 40–270 specimens/g of sediment) is noted in the layer of 200–204 cm. This is accompanied by a similar decrease in the $CaCO_3$ and C_{org} content in sediments suggesting a general decline in the produc-

tivity of the surface water layer that may be associated with a decrease in temperature after the first phase of glacial thawing. In the oxygen-isotope variation curve for *N. pachyderma* shells, the interval in question is marked by highest $\delta^{18}O$ values (about +3.5‰) as compared to others characterizing the deglaciation period (Fig. 3). The respective temperature fall in surface water is also evident from a reduced (several times lower) abundance of subpolar *G. bulloides* and a simultaneous, nearly total disappearance of relatively thermophilic *N. pachyderma* dex. and *G. quinqueloba* from the assemblage. Foraminiferal shells are poorly preserved. They are damaged, fragile, corroded, and often represented by fragments. The poor preservation suggests that the decreased shell abundance and reduced $CaCO_3$ content in sediments can be results of lower productivity and, largely, of intensified dissolution of carbonates at the sea floor. The accumulation time of sediments under consideration (their age is about 10800 years, Gorbarenko *et al.*, 1998) corresponds to the Younger Dryas event, though other researchers reported somewhat different age determinations. This climatic event is recorded in many regions of the northern hemisphere, including the North Pacific in particular (Keigwin, 1987; Keigwin *et al.*, 1992; Kennett, 1995; Kennet and Ingram, 1995; Kennet and Venz, 1995; Zahn *et al.*, 1991; etc.), and also in New Zealand (Broecker, 1997).

The interval of 195–160 cm characterized by smooth diminishing of $\delta^{18}O$ values corresponds to the second deglaciation phase or Termination 1B (Gorbarenko *et al.*, 1998) and has the highest absolute abundance of foraminiferal shells (up to 4600 specimes/g), predominantly of *N. pachyderma* sin. and *G. bulloides*. The maximum content of $CaCO_3$ and C_{org} in the sediment is 9.2% and 1.5%, respectively. The layer of 184–174 cm evidently corresponds to the peak time of the event 9230–8770 years ago.

At the end of the event (the layer of 172–160 cm), the abundance of foraminifers and $CaCO_3$ content were reduced, and simultaneously the amount of shell fragments notably grew. This suggests a high corrosive activity of water at that time. Calcium carbonate dissolution at the sea floor was more intense in the second than in the first phase of ice thawing, particularly, at the end of the former about 8000 years ago. The growing intensity of foraminiferal shell dissolution during the two phases of ice thawing might be a result of the progressive rising of surface water bioproductivity, and, consequently, of higher concentrations of organic matter in sediments.

On the whole, the section interval of 223–160 cm comprising two deglaciation phases revealed a higher abundance of planktonic foraminiferal shells than glacial sediments. Variable concentrations of planktonic foraminifers in glacial and deglacial deposits characterizing a chiefly terrigenous type of sedimentation reflect a changing productivity of surface water rather than the

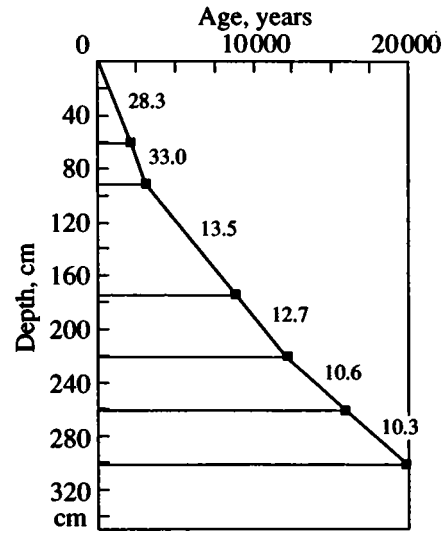

Fig. 4. Changes of average sedimentation rates in the investigated area for the last 20000 years.

effect of shell dissolution at the sea floor; the last factor only contorted somewhat the general trend. The greater abundance of foraminiferal shells in deglaciation sediments as compared with that in glaciation deposits and a certain increase in the average sedimentation rate from 10.5 to 13.5 cm/k.y., respectively, i.e., the greater effect of dilution factor, also suggest that the foraminiferal productivity in surface waters was the main factor that controlled their abundance in sediments (Fig. 4).

The planktonic foraminiferal assemblage from the postdeglaciation ooze in the upper part of the section (160–0 cm) is less rich in subpolar *G. bulloides*, representing from 10–15 to 2–8% of the total amount of foraminiferal shells, and *N. pachyderma* sin is predominant here.

The highest concentration of shells is recorded in the interval of 132–120 cm, where they are mostly represented by *N. pachyderma* sin. (the peak abundance is established only for this species). Maximum $CaCO_3$ concentration (7–9%) is measured in the interval of 135–103 cm and remains high even if the total abundance of foraminiferal shells in sediments of the upper part of the interval decreases.

The planktonic foraminiferal assemblage from diatom ooze in the 110–0 cm interval characterizes the last 4000 years and is distinct because of a low *G. bulloides* content (4–15%) and abundance maximum of *G. quinqueloba* relative to the other interval of the studied section (from 1–3 to 4–15%). The cryophilic affinity of the assemblage is reflected in the highest relative concentrations of *N. pachyderma* sin. The assemblage is characterized by lowest abundance values of planktonic foraminifers as compared with underlying terrigenous and transitional silts of the last glaciation and deglaciation. The shell walls are fragile, and many specimens are partly dissolved. The amount of intact shells here is

either close to that observed in terrigenous glacial ooze, or several times less. The low abundance of planktonic foraminifers certainly marks the intensification of dissolution processes at the sea floor at that time. This is also evident from the fact that the sedimentation rate was two–three times greater at this level than during glaciation and deglaciation time (27–33 cm/k.y.), but the $CaCO_3$ content in sediments remained high (5–8%). A sharp growth of siliceous microplankton productivity in surface waters brings about the sedimentation rate increase. The amorphous SiO_2 content in sediments of the interval of 110–0 cm ranges from 40 to 55%; C_{org} is between 0.5 and 1.4% (Figs. 2, 3).

RESULTS AND DISCUSSION

The revealed changes in the taxonoimic composition of planktonic foraminifers and isotopic composition of their shells from the core section V34-90, taken from the eastern slope of the Academii Nauk Rise in the Sea of Okhotsk, enabled us to detect major climatic events of the last 20000 years: glaciation (20000–12500 years ago), deglaciation (12500–8000 years ago) with a well-expressed cooling episode (Younger Dryas, about 10800 years ago), and postglaciation (the last 8000 years).

During the glaciation (about 20000–12500 years ago), surface waters of the Sea of Okhotsk were inhabited by an impoverished assemblage of planktonic foraminifers represented by *N. pachyderma* sin. most abundant in the assemblage (71–93%), *G. bulloides, G. quinqueloba*, and dextral *N. pachyderma*. In the interval of about 15 000–14 000 years, the content of subpolar *G. bulloides* species is two–three times lower, and *N. pachyderma* dex. completely disappears, which should be a result of the glaciation peak. This is also confirmed by high $\delta^{18}O$ values here 3.5‰. A somewhat increased relative content of these species (to 31 and 7% respectively), is recorded in the interval of 270–273 cm and seems to indicate a certain warming, which, however, is not reflected in the isotopic curve (Fig. 3).

A good preservation of planktonic foraminifers and insignificant quantity of their fragments in the glaciation sediments, along with their relatively low concentrations and minimum $CaCO_3$ content, as compared to other intervals of the section, suggest that the carbonate dissolution at the sea floor was hardly significant at that time.

A better preservation of carbonate shells of planktonic foraminifers in glacial sediments relative to the postglacial period in the study area apparently was associated with different chemical properties of bottom water. An increased $CaCO_3$ content in oceanic sediments of the Pleistocene glacial epochs and its decreasing in interglacial periods was noted long ago (Arrhenius, 1952). However, causes of such variations are still discussed. At present, the majority of researchers are of opinion that variable concentrations of calcium carbonate in sediments are mainly controlled by changing properties of bottom waters, whereas changes in the calcareous plankton productivity in surface waters play a secondary, though important role (Chuey *et al.*, 1987; Hovan and Rea, 1991; Keigwin *et al.*, 1992).

The recent warm deepwater current from the Krusenstern Strait is mainly directed to the northeastern region of the Sea of Okhotsk. In the area of core section, the warm (1.8–2.3°C) Pacific water mass has a minimum oxygen concentration (2.0–2.1 ml/l) and is chemically aggressive with respect to calcium carbonate.

Recent studies revealed that the geological regime in North Pacific varied during the late glaciation and deglaciation, though the changes were not as great as in the North Atlantic (Zahn *et al.*, 1991).

As some researchers reported (Keigwin, 1987; Zahn, 1991), no deep-sea water poor in biogenic substance was formed during the last glacial epoch in the northern part of the Pacific Ocean, as it was characteristic of the North Atlantic. At the same time, certain data available suggest that intermediate water masses poor in biogenic elements and rich in oxygen may exist in the Pacific within the depth interval of 700–2600 m. These waters may originate somewhere in the northern part of the Pacific Ocean (Duplessy *et al.*, 1988) and penetrate into the Sea of Okhotsk from the ocean through straits, which are as deep as 2000 m, i.e., approaching the depth interval of intermediate waters; this phenomenon could be responsible for a better preservation of calcareous shells during the glaciation. There is another point of view suggesting that, during the last glacial epoch, the Sea of Okhotsk itself could serve as source area of the intermediate water mass in the North Pacific (Zahn *et al.*, 1991).

The deglaciation period (about 12500–8000 years ago) marks the rearrangement of the hydrologic regime in the Sea of Okhotsk that affected both the surface and bottom waters. At the beginning of the deglaciation, the total abundance and diversity of species in the foraminiferal community increased; assemblages of this level usually include more thermophilic species as compared with those of the glacial time. This is evident from a higher abundance of subpolar *Globigerina bulloides* and appearance of *Globigerinita uvula, G. glutinata*, and *Globorotalia scitula* forms not encountered in glacial sediments. In the section interval corresponding to the deglaciation period, there are two peaks of the shell and $CaCO_3$ concentrations, which coincide with negative excursions of $\delta^{18}O$ values and agree with two deglaciation phases T1A North Pacific and T1B North Pacific (Gorbarenko *et al.*, 1998).

Accumulation rates of terrigenous pelitic ooze during the deglaciation were close to those during the glaciation (13.0 and 10.5 cm/k.y., respectively). This indicates that there were no significant changes in the degree of the shell material dilution with terrigenous components during the deglaciation period.

At the beginning of deglaciation, the dissolution of calcium carbonate at the sea floor intensified, as it follows from the increased abundance of shell fragments. Throughout the transitional period, the dissolution of carbonate shells was much more intensive than during glaciation. The amount of intact shells in the deglaciation sediments is not more than 2–3 times greater than their abundance during the glacial time, whereas the $CaCO_3$ content becames 4–5 times higher due to increased occurrence of shell fragments (Figs. 3, 4). This shows that bottom waters of the Sea of Okhotsk were at that time more aggressive with respect to calcium carbonate.

Variations in the abundance of planktonic foraminifers and $CaCO_3$ content in sediments could be associated with several factors, such as changes in the calcareous plankton productivity in surface waters, intensity of dissolution at the sea floor, and sedimentation rate. Taking into consideration that sedimentation rates during the deglaciation were higher than in the glaciation time, and $CaCO_3$ content in sediments became many times greater despite the intensive dissolution, the grown abundance of planktonic foraminifers can be attributed to the increased productivity of surface waters. This is also evident from a variable C_{org} content in sediments. The lowest productivity of planktonic foraminifers during the deglaciation was recorded in the Younger Dryas, when both the total abundance of shells and $CaCO_3$ content in sediments simultaneously declined.

The appearance of *Globigerinita uvula* and *Globorotalia scitula* at the beginning of deglaciation (12180 years ago) and some time later of *Globigerinita glutinata* may be explained by their penetration at that time through the straits with the Oyashio current from the Pacific Ocean. In glacial sediments, these species are missing, even from layers with a high total abundance of species; therefore, their absence can hardly be attributed to the effect of dissolution. During the glacial time, the northern boundaries of distribution areas of these species must be extending into the ocean southward of the East Kamchatka and Oyashio currents in the northwestern Pacific Ocean. Data on sections in the subarctic Pacific region suggest that, during the last glaciation, the subarctic gyre-circulation system, the Oyashio current included, differed notably from the recent one, and the boundaries of distribution areas of relatively thermophilic planktonic foraminifers were shifted southward (Thompson and Shackleton, 1980). At present, *G. uvula* is widespread in the subarctic Pacific displaying the maximum concentration in its northwestern part near the Kurile Islands. *G. glutinata* is also present in this region. At the beginning of the deglaciation, the distribution limits of both species migrated evidently to the north, and they penetrated together with ocean waters into the Sea of Okhotsk.

At present, *G. scitula* is known only from the high north latitudes of the northeastern Pacific (Bradshaw, 1959). There is no published data pointing to its occurrence in the northwestern Pacific. Therefore, we can not firmly assume that the species penetrated into the Sea of Okhotsk from the Pacific Ocean through straits of Kurile Islands at the beginning of the deglaciation along with two other species. Probably, the occurrence of *G. scitula* in this sea is related to its migration from the Sea of Japan, when both basins were connected through the Soya Strait 13180 years ago (Belyaeva and Burmistrova, 1997), or maybe 12180 years ago, if we take into consideration the reservoir-corrected reduction of age determinations.

The transformation of the hydrologic regime in the Sea of Okhotsk continued during the postglacial period (the last 8000 years) and finally resulted in formation of the modern structure of its water mass. On the whole, the composition of planktonic foraminifers did not experience substantial changes as compared with this fauna of deglaciation period. As before, the assemblage is dominated by the sinistral *Neogloboquadrina pachyderma* variety associated with less abundant *Globigerina bulloides* and *Globigerinita quinqueloba*, and with scarce *G. uvula, G. glutinata, Globorotalia scitula*, and dextral *N. pachyderma* (table).

The beginning of the last period is marked by peak abundance of planktonic foraminifers, high $CaCO_3$ content in sediments, and an insignificant negative excursion of $\delta^{18}O$ (0.5‰). For the time of 7000–5000 years ago, this peak coincides with a weakly expressed isotopic event T1C North Pacific (Gorbarenko *et al.*, 1998) and with the climatic optimum revealed by palynological data in Hokkaido (Tsukada, 1988) and at site V32-161 in sediments of the Akademii Nauk Rise in the Sea of Okhotsk (Morley *et al.*, 1991).

In the studied interval of the 142–120 cm corresponding to that event, the relative abundance of subpolar *Globigerina bulloides* sharply (2–3 times) decreased and was as low in the following period just up to the present. At the same time, dextral varieties of *N. pachyderma* virtually disappears completely, though it was persistently present before in small quantities in the glaciation and deglaciation sediments. In view of that, the assemblage of planktonic foraminifers became more cryophilic as compared to those of the previous glaciation and deglaciation periods. This seems to be a consequence of the specific features of the surface water mass structure in the Sea of Okhotsk, which is supposed to be similar to the recent one (Morley *et al.*, 1991).

Comparison of the oxygen isotope parameters of carbonate in *N. pachyderma* sin. and *G. bulloides* shells and their current distribution areas in the subpolar North Pacific (Reynolds and Thunell, 1985; Sauter and Thunell, 1989) suggests that the cryophilic affinity of planktonic foraminiferal assemblage that was impoverished in *G. bulloides* during the last 8000 years resulted from the temperature minimum, which appeared in the Sea of Okhotsk. Values of $\delta^{18}O$ in *G. bulloides* shells

are by 0.5‰ lesser than in *N. pachyderma* sin. Such a difference suggests that the bulk mass of shell carbonates of these species was accumulated under different conditions. Accordingly, the bulk mass of calcium carbonate of *G. bulloides* originated in warmer and less saline water than that of *N. pachyderma.*

As is established by Reynolds and Thunell (1985), May–early June is a time of *G. bulloides* highest productivity in the high latitudes of the northeastern Pacific. Thus, the reproduction period of this species in the Sea of Okhotsk coincides with the very beginning of summer warming of surface waters and with the appearance of their thermal stratification and subdivision into the surface and deeper (Sea of Okhotsk) water masses. This process reaches its peak in August, when there appeared the shallow-water temperature minimum along with the high density and temperature gradients. Surface waters become sufficiently warmed up only to the depth of about 30 m, whereas the temperature at deeper levels remains close to that of the winter period (about 0°C or lower). Studies on the plankton distribution in the water column showed that the planktonic biomass sharply decreases at the depth below 25 m (Moroshkin, 1966). Evidently, the aforementioned seasonal temperature fluctuations within the surface water layer of the Sea of Okhotsk, which are disagreeable to the period of maximal reproduction activity of *G. bulloides* (and may be of *N. pachyderma* dex. as well), unfavorably influence the evolution of planktonic foraminifers in this region and, consequently, they could be responsible for the decreased relative abundance of the species in sediments of the postglacial period.

In contrast to the recent water mass structure presumably characteristic of the entire postglacial period, the surface water mass of the glaciation and postglaciation time must have had a lower spring–summer temperature. This favored the vertical mixing of waters and might cause a decline or even disappearance of the temperature minimum. Accordingly, conditions for development of *G. bulloides* and *N. pachyderma* dex. were more favorable.

At the same time, the above factors do not have any effect on the development of such species as *N. pachyderma* sin. and *G. quinqueloba*, both commonly confined to cold well-mixed waters. Investigations carried out in the Pacific high latitudes (Reynolds and Thunell, 1985) showed that their maximum productivity corresponds to the spring time (April–May), when surface waters are coldest and isothermal.

Though the abundance of planktonic foraminiferal shells in the postglaciation sediments is diminished relative to those of deglaciation and even glaciation time (Fig. 2), the plankton productivity in the latest period apparently was sufficiently high, as it follows from the high $CaCO_3$ content in sediments. The reduced abundance of shells appears to be primarily associated with an intense shell dissolution and a sharp rise in the sedimentation rate at that time due to biogenic silica influx to the floor, which caused mechanical dilution of foraminiferal shell concentration.

The recent structure of water masses in the Sea of Okhotsk existed for the last 4000 years. This period was characterized by the constant high productivity of siliceous phytoplankton and, consequently, by the highest content of amorphous silica in sediments, which ranges in the section from 30 to 40%. The planktonic foraminiferal assemblage of that period has the greatest relative abundance of *G. quinqueloba* and *N. pachyderma* sin. and is marked by the almost complete absence of warm-water species *G. scitula* and *G. glutinata*. A poor preservation of foraminiferal shells frequently fragmented suggest that the dissolution at that time was most intense for the entire considered history of the Sea of Okhotsk. This might be a consequence of the specific hydrological regime at the sea floor, and also could result from an elevated content of organic matter in the postglacial sediments, which accumulated organic remains. The oxidation of the latter lowers concentrations of oxygen dissolved in the water, thus intensifying the corrosive influence on calcium carbonate.

Postglacial hydrologic changes in the Sea of Okhotsk affected the character and sedimentation rate on the whole. During the considered 20000 years, the average sedimentation rate in the study area persistently went up, first rising gradually from 10.5 cm/k.y. to 13.5 cm/k.y. during the glaciation and then, during the deglaciation, up to 14.5 cm/k.y. at the beginning of postglacial time; finally, during the last 4000 years, the sedimentation rate ranged between 28.3 cm/k.y. and 33 cm/k.y. (Fig. 4). In the glaciation and deglaciation periods, sediments were formed as a result of influx of terrigenous, chiefly clayey material. The total amount of biogenic carbonate and siliceous material at these levels is no more than 10–15%. In the postglacial period, beginning from the age boundary of about 6000 years, the sedimentation in the Sea of Okhotsk sharply changed due to the mass flux of biogenic silica to the sea floor. The SiO_2 content in sediments of that time (interval of 0–130 cm) ranges from 25 to 50%. As the content of biogenic (carbonate and siliceous) material in the studied section rose, parallel to the growing sedimentation rate, from less than 10% in the glacial sediments to 20% during the deglaciation time, and to 55% and more in the postglaciation period, the terrigenous influx during the entire period under consideration did not change significantly.

CONCLUSION

The analysis of quantitative and qualitative changes in planktonic foraminiferal assemblages from the Late Pleistocene–Holocene sediments of the Sea of Okhotsk, data on isotopic composition of their shells, and results of radiocarbon datings allowed us to follow the history of climatic and hydrologic changes during

the last 20000 years. The succession of events is supposed to be as follows:

(1) In the period of 20000–12500 years marked by the highest $\delta^{18}O$ values and corresponding to the glaciation event, the Sea of Okhotsk was inhabited by the cold-resistant assemblage of planktonic foraminifers represented only by three species (dextral and sinistral *Neogloboquadrina pachyderma, Globigerina bulloides*, and *Globigerinita quinqueloba*). The interval of 14000–15000 years was characterized by the lowest temperature of surface waters. The decreased abundance of foraminifers and minimal content of $CaCO_3$ and C_{org} in sediments of this interval suggest the lowest productivity of calcareous and siliceous plankton for the whole history under consideration. A good preservation of foraminiferal shells and low $CaCO_3$ content are indicative of weak intensity of shell dissolution.

(2) During deglaciation (12 500–8000 years), the planktonic foraminiferal assemblage was marked by the maximum species diversity (seven species) because of the appearance of relatively thermophilic species *Globorotalia scitula, Globigerinita uvula*, and *G. glutinata*. They could arrive either from the Pacific Ocean through the Kurile straits or from the Sea of Japan through the Soya Strait that had been opened about 12180 years ago (Belyaeva and Burmistrova, 1997). Two negative excursions of $\delta^{18}O$ values that are clearly distinguished on the isotope curve correspond to warming phases (T1A North Pacific and T1 B North Pacific episodes) also reflected in the increased abundance of foraminiferal shells and higher $CaCO_3$ content in sediments. These episodes are separated by a positive excursion of $\delta^{18}O$ values by 1‰, which coincides in time with the Younger Dryas episode about 10800 years ago (Gorbarenko *et al.*, 1998). The beginning of deglaciation is also characterized by a notable intensification of shell dissolution indicating that the Sea of Okhotsk basin was filled with "young" water originating either in an intermediate layer in North Pacific (Duplessy *et al.*, 1988) or, probably, in the Sea of Okhotsk itself (Zahn *et al.*, 1991).

(3) During the postdeglacial time (the last 8000 years), the transformation of the hydrologic regime of the Sea of Okhotsk continued. It started during the deglaciation and 4000 years ago led to the creation of the present-day structure of water masses with a clear spring–summer stratification of surface waters and well-defined temperature- minimum. This caused a rapid growth of siliceous plankton productivity and an intence flux of biogenic silica to the floor that resulted in the twofold increase in the sedimentation rate. The planktonic foraminiferal assemblage of this time was relatively cold-resistant as is evident from the predominance of *N. pachyderma* sin. and the reduced significance of *G. bulloides*. The decreased relative abundance of *G. bulloides*, the dweller of near surface waters, was, apparently, a consequence of the time difference between the reproduction period of the species and the seasonal warming of surface waters in the Sea of Okhotsk. The first half of the postglacial period (about 6000 years ago) is marked by the well-expressed abundance peak of planktonic foraminifers and negative excursion of $\delta^{18}O$ values, concurrent to an isotopic event revealed here earlier (Gorbarenko *et al.*, 1998), and to the climatic optimum established in Hokkaido on the basis of palynological data (Tsukada, 1988) and in the Sea of Okhotsk (Morley *et al.*, 1991). Calcium carbonate dissolution at the sea floor continued in the postglacial period and was at its maximum during the last 4000 years, when it resulted in the poor preservation of planktonic foraminifers, though with the retained high productivity in the surface layer and high $CaCO_3$ content in sediments.

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Stratigraphic Code or Code of Stratigraphic Nomenclature?

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The paper continues the discussion of typification problems considered in my previous article (Lazarev, 1997). Suggestions to reform the typification procedure (the only possible way to stabilize the stratigraphic nomenclature) were met by my opponents as something absolutely new and unusual. Such a reaction was particularly unexpected from stratigraphers and paleontologists, who are well acquainted with the method used to stabilize nomenclature in biological classification. References to the specifics of different sciences are not pertinent here: my discussion was concerned with the question of what is the nomenclature type in general, and why it cannot be combined with the substantial type (archetype). In my previous paper, I summarized well-known views on how the priority principle can be competent, and why this becomes unnecessary under the supremacy of the chronostratigraphic paradigm. By the way, views on the relationships between classification and typification in stratigraphy were most clearly expressed in the books of V.A. Krasilov (1977, 1985).

It is also surprising that problems related to the usage of the multivalent term "type" remain abstruse for many paleontologists and stratigraphers. It should be emphasized that almost four decades have already passed since the time when the International Code of Zoological Nomenclature (ICZN) appeared, thus ending the period of discussion and doubt, and when the concept of the nomenclature type as a basis of the nomenclature stability was separated from the general concept of "type." This was in biology, but not in geology, where the concept of type remains at the same level as it was in the 1950s. Unfortunately, the last decades were marked not only by a certain stagnation, but also by destruction of stratigraphic classification. At present, the method of construction of the stratigraphic scale is limited by the selection of GSSP (actualy the points of lower boundaries of stratigraphic units), which has almost nothing to do with the classification procedure. The disappearance of classification made typification (*sensu stricto*) superfluous. That is why most consistent adherents of chronostratigraphy rejected the idea of the stratigraphic unit stratotype (Remane *et al.*, 1997), which was easy to predict (Lazarev, 1997).

Below, I analyze the methodological metamorphosis, which remained imperceptible for most stratigraphers and which makes it impossible to create a true nomenclature code.

REGULATION CONCERNS SUBSTANTIAL PROBLEMS, WHEREAS JURISDICTION, ONLY TECHNICAL (SUBSIDIARY) ONES

To start with, two spectra of problems should be defined. The first of them unites the following substantial problems: (1) general (theoretical) aspects of stratigraphy; (2) construction of stratigraphic scales; (3) stratigraphic terminology (but not nomenclature!); (4) correlation of stratigraphic units and identification of their ranges and boundaries. Even most general of these problems is subjected to regulation (via glossaries, manuals, and so on). Nonetheless, jurisdiction, i.e., very strict prescriptions which are to be strictly followed by all the stratigraphers regardless of their theoretical orientation, is contraindicated to all these problems. The second spectrum of problems includes those which do not concern directly substantial aspects of stratigraphy, i.e., the problems of nomenclature and, most important, the selection of competitive names of stratigraphic units. Only subsidiary problems of such a type can be subjected not only to regulation, but also to strict jurisdiction without adverse effects to science.

Regulation of substantial problems is also manifested in collective (sometimes, international) decisions and agreements. Problems of three levels (the generality degree) can be conventionally distinguished: (1) those of general theoretical character concerning, for instance, principles of stratigraphy or the nature of stratigraphic scales; (2) stratigraphic methods, for instance, those related to the ranking of particular stratigraphic features during subdivision and correlation of stratigraphic units of a certain scale (general, regional, or local); (3) selection of a particular boundary between stratigraphic units, which is necessary for compilation of legends for geological maps. Problems of term and concept definition can be referred to any of these levels depending on the degree of their generality. I should emphasize that all three levels of the generality are interrelated, therefore, even the third level, closest

to the practice, can significantly depend on the solution of problems at the first level. Thus, if we consider the general stratigraphic scale only as representing a general (conventional) rule of measuring time which does not depend significantly on the problem of boundary ranking, then we should consider, when selecting the particular boundary of the stage (this work is now conducted in the frame of international working groups), first of all a demarcation level close to the previously accepted boundary. If we consider the stratigraphic scale as a scheme of substantial classification, then problems of hierarchy of stratigraphic units are not indifferent for us in the course of revision of stratigraphic scales, and the selected position of a particular boundary of the stage can differ from that selected in the first case. Most important is that all three generality levels always concern substantial problems, and, therefore, the demarcation line separating the jurisdiction procedure (the application sphere of the Code) passes between, but not within, all of them and the formal nomenclature problems. This contradicts, to a certain extent, the viewpoint of S.V. Meyen (1989), who always used the term "regulation," not "jurisdiction," when discussing the same point, so the demarcation line between these concepts was not outlined by this researcher. Thus, we will discriminate the concept of regulation (agreement or restriction, which touches upon substantial problems but is of a temporary nature and changes after transformation of the content proper) and the concept of jurisdiction (strict rules and restrictions, which have a status of law; they are rated at the indefinitely long usage and touch only upon formal nomenclature problems).

CODES (JURISDICTION): NOMENCLATURE—YES, INTENSION—NO!

Hence, any substantial problem should not be principally subjected to jurisdiction, and, consequently, the Code, as a juridical document, should not touch upon these concepts. These undoubted statements are met with incomprehension by stratigraphers, because the situation was developed in such a way that the type appeared to be connected in stratigraphy both with the name and content. Nowadays, we have a triune, closely interrelated system "type–name–content," according to which any change in the stratigraphic unit meaning inevitably results in its name transformation. So far, only one way out situation has been found—the bridling of meaning. This led stratigraphers to equalization of regulation and jurisdiction, as well as to unprecedented striving for conservation of meaning, as it is evident from substitution of reasonable regulation of subtantial notions and terms (via glossaries, manuals, and so on) by jurisdiction (via codes). Meanwhile, our language, including the science language, will always remain polymorphous, because one cannot define the unambiguous connection between the concept expressed by words and the essence, and, therefore, terms should be explained rather than strictly defined (Nalimov, 1979). Figuratively speaking, nobody can make a swallow live the mole's life: this is only the unfortunate episode of a fairy tale. In practice, it is important to find the golden mean between the clearness of statement and a quite loose understanding of the relevant concept every time when formulating the term definition: we are able to understand a word (term) in the particular context of other words.

Regulation of substantial concepts on the basis of different manuals and agreements is justified only by tactical considerations: any connotation in any science cannot remain unchangeable. Another thing altogether are documents of a juridical nature (codes in the science) rated at the indefinitely long-term (strategic) perspective, but not at the certain scientific paradigm; the codes should not touch upon substantial problems. Biological codes regulating exclusively nomenclature problems are of this sort; they will basically remain unchanged during an indefinitely long time, as long as biologists will use the hierarchical form of systems (Lazarev, 1997).

From the "Stratigraphic code," it is clear that its authors used the term "code" as something intermediate between instruction and manual. More important is that the notion "nomenclature" is perverted in this document (as well as in the codes of other countries). There are sections dedicated to the nomenclature and even a chapter "The Priority" that is the basis of any nomenclature. In fact, nomenclature is suppressed by the content in this and other "codes," and the priority rule cannot be used because the name priority is mixed with the intension priority. The close tandem "name–content" became psychologically indivisible for stratigraphers in last decades; I always hear the same argument: "otherwise nobody understands anything" (for more details, see below).

All "codes" available now in stratigraphy are helpless exactly in the solution of nomenclature problems, and their titles lack the word "nomenclature." The only exception is the first code *Code of Stratigraphic Nomenclatue* published in the United States. Unfortunately, its title in no way corresponded to the content.

Only code aimed exclusively at the solution of nomenclature problems can become a universal juridical document. Elaboration of such a code is quite possible, and its name "International Code of Stratigraphic Nomenclature" would reflect its content.

TWO MODELS OF THE STRATIGRAPHIC SCALE ESSENCE: "SCALE AS A RULE OF MEASURING TIME" OR "SCALE OF THE INTENSION CLASSIFICATION"

My habit of classification causes me to categorize main approaches in stratigraphy. The opinion that there are only two stratigraphies, good (competent) and bad (incompetent), is be correct when the particular prac-

tice of sequence subdivision and correlation is kept in mind. The same pragmatic opinion is, however, incorrect in a strategic perspective, i.e., when it concerns most general concepts, for instance, the intension of stratigraphic scales. Indeed, the sentence about two stratigraphies in this case would be as follows: there are two approaches to stratigraphy—good (mine) and bad (of my opponents).

In fact, the pragmatic approach to the problem of creating stratigraphic scales can be expressed in the following way: stratigraphers have never had a consensus of opinion on how to create their scales, and boundaries of stratigraphic units constantly changed their position (their hierarchy changed); is it not better to have a stable (standard) base for time ordering of geological records? This approach is embodied in the chronostratigraphic paradigm predominant nowadays (see below). The opposite paradigm merits the status of the traditional one because it retains the privilege of the scale for developing the intension.

In some sense, contraposition of two paradigms in the modern stratigraphy can be related to the contraposition of well-known notions such as absolute and relative time, European and American stratigraphic scales, typological (focused-on-sequence) and population (systemic) approaches. Most significant of these contrapositions is the attitude to the notion of time in geology, although the essence of this notion is too complicated. In this connection, the books by K.V. Simakov (last published in 1996) are remarkable and very interesting with respect to references cited in abundance but naive in the attempt to solve one of the basic problems of being—the problem of time. The author showed the most peculiar feature of the chronostratigraphic paradigm, which he himself subjects to criticism, the striving for introduction of strict and ultimate statements into science, without which, he believes, true classification is impossible.

It is easier to classify ideas than their carriers because of the ideological eclecticism of most stratigraphers; this is particularly clear from both editions of Russian codes. The most distinct demarcation line, which separates adherents of two paradigms of the modern stratigraphy, is, in my opinion, the attitude to the essence of stratigraphic scales, i.e., to one of two scale models: the scale aimed at the intension classification, or the scale with fixed points usable as a rule of time measurement. Therefore, it is better to term the traditional model of the scale as classificatory, whereas another one, accepted now as official, corresponds to the point model. Ecology and causality are very useful when GSSP is selected; unacceptable in the chronostratigraphic methodology is an alternative procedure of GSSP conservation in the particular section, which means, in perspective, the complete separation of the selected point from the inevitably changeable intension. The philosophy of point stratigraphy was unequivocally expressed already in the first edition of *"Guidelines..."* (Lazarev, 1997): our interpretation of boundaries is subjective and changeable, whereas the particular point in the particular section is objective and absolute. After fixation of all GSSPs, when this "happy" future is attained, their general succession can be numbered, but the hierarchic form of the scale (an illusion of classification) will remain a sort of atavism inherited from the last developmental stage of the more traditional stratigraphy.

It was suggested (Zubakov, 1992) to use simultaneously, as a compromise variant, both models of stratigraphic scales: the existing model, which becomes progressively more conventional, i.e., the point model, and the classificatory (intension) model, which is termed as the nomotetic one. The inadequacy of this approach was discussed earlier (Lazarev, 1997).

CODES AND CLASSIFICATION: THE PROBLEM OF SELECTING THE ANALOGIES

If the stratigraphic scale is considered to represent a scheme of intension classification, then its right for the intension development should be admitted, through this contradicts the opinion of stratigraphers striving for scale stability. The method of analogies (comparative method) is universal in science and should help to find an optimal solution of the problem. Indeed, it is desirable to find the closest analogies. This procedure already includes elements of classification, as does any other scientific analysis. In addition, classification (*sensu stricto*) implies an organizing of data into a certain form. There are three types of classification: combinative, parametric, and hierarchic. The first two classifications are peculiar of exact sciences, whereas the last one is usually used in historical (by their nature) sciences, because processes directional and irreversible in time are probably more important in this case than cyclic ones.

Simakov (1997) believes that an analogy of stratigraphic classification with that of chemical elements and substances is more appropriate and doubts the validity of my analogies, asking what is indicative of the water type—oxygen or hydrogen? It is easy to note that the first classification deals with objects of the same level (stratigraphic units of a variable rank), while the second one, with objects of two levels (chemical elements and substances). Indeed, it would be unreasonable to propose substances, for instance, $CaCO_3$ or SiO_2, to represent types of stratigraphic units. Simakov's analogies are also inappropriate because of formal reasons. In classifications peculiar of exact sciences, there is no problem of selection of competitive names, which represents a central problem of typification considered in my article. Other statements of Simakov's criticism are also not quite correct: my arguments to substantiate the necessity of a common approach to the nomenclature (technical) aspect of any classification and relevant arguments to support the separation of intension problems (the sphere of classi-

fication) from nomenclature problems (the sphere of typification) were interpreted by him in such a manner as if I propose the principles of biological classification to be introduced into stratigraphy and object to the use of definitions of intension in stratigraphy.

The problem of selection of competitive names is common and makes reasonable an analogy between biological and stratigraphic classifications; moreover in this case, one can speak about identity, not analogy, in the principal solution of the same problem, which is of a subsidiary (technical) nature and only indirectly concerns the substantial core of the problem. Specifics of stratigraphy will strongly affect only particular phrasing of the code clauses, but specifics should never touch upon principles of the code nomenclature: (1) nomenclature and intension problems should strictly be separated, (2) type is a standard of the name, but not of the content, (3) the hierarchical principle is a constituent of the typification procedure.

The fundamentally different views on the problem of analogies are shared by most consistent adherents of chronostratigraphy, who refuse to speak at all about analogies between the stratigraphic scale and any other classification. Indeed, during the last decades, the development logic of chronostratigraphy resulted in the destruction of the stratigraphic classification: "the scale as the intension classification" turned into "the scale as a rule of measuring time." This allows adepts of chronostratigraphy to rightly reject any comparison between scales and classifications and to state that "these are absolutely different things." Owing to this logic, time appears, obviously or unobviously, to be torn off the historical–geological processes. A stratigraphic unit for chronostratigraphers is an object that was formed during a certain time period and recorded in the stratotype section and/or with a help of limitotypes (*Stratigraficheskii kodeks*, 1992, p. 25). It is apparent that differences in opinion are of a more general methodological nature. Therefore, the next two sections will be dedicated to the analysis of the development and essence of relatively new methodology in stratigraphy.

CREEPING REVOLUTION OF THE 1960s: TRANSFORMATION OF THE METHOD OF THE GENERAL STRATIGRAPHIC SCALE CONSTRUCTION

The early history (prior to the 1960s) of the establishment and evolution of stratigraphic units of the general stratigraphic scale was excellently reviewed in the two-volume monograph by G.P. Leonov (1973, 1974). Following the history of the study of Paleozoic systems, beginning with the Cambrian, one can see that their conception and ranges significantly changed. The same is true for all stages of the general scale. Probably, the most demonstrative is the case of the conception and range of the Silurian System. Initially, it was defined by R. Murchison as comprising six formations from Llandeilo to Ludlow; moreover, the range of the Silurian (more exactly, its lower boundary) was determined using the criterion of "fossil-bearing potential" of the relevant deposits. Later, as fossils were discovered in successively lower layers, the author of the system himself displaced the Silurian boundary down the sequence to the Lower Cambrian (in its modern range). The name "Silurian" was retained despite this significant change of the system range. From the recent point of view, Murchison's idea to conserve the name "Silurian" could be somehow excused, because only the extension (range) of the stratigraphic unit, not its intension ("fossil-bearing potential"), has been changed. Moreover, the name "Silurian" is authorized even at present, although a part of this system is categorized as the autonomous Ordovician System. The last decision was, among other things, a compromise between adherents of Murchison and Sedgwick, who differ in opinion in regard to the conception and range of the Cambrian and Silurian systems. This compromise was accompanied by both a substantial change in the extension of these systems and the introduction of a new intension—the difference in fauna composition that grounded the fundamentally new three-member subdivision of the Cambrian–Silurian deposits (three successive faunas).

A similar situation was characteristic of the history of the Carboniferous System, which primarily was accepted as significantly wider in range, spanning the Devonian System and the lower part of the Permian System. Moreover, the Permian System was also defined by Murchison as beginning with the Kungurian deposits. Stages established in the last century also experienced significant modifications. For instance, the Artian Stage was primarily sufficiently wider in range and included Sakmarian and Asselian stages of the recent scale. Nonetheless, the name "Artian Stage" is in use today.

Thus, the conception and range of stratigraphic units of the general scale significantly changed as relevant deposits were better studied. This is a normal consequence of the progress of knowledge. Two points are important here. First, changes in the range of stratigraphic units were connected with changes in the hierarchy of their boundaries, and this was related, in one way or another, to the classification process during the general scale elaboration. Second, it is important that, up to the middle of our century, stratigraphers (as well as biologists), who had no strict juridical basis for the nomenclature, spontaneously used the priority law, which allowed them to keep initial names of stratigraphic units in any case of boundary displacements. Indeed, nobody conceived the idea of rejecting the name "Silurian" after the significant displacement of its lower boundary, or after separation of the autonomous Ordovician System. Similarly, nobody objected against the usage of the name "Artian" after the separation of two autonomous Sakmarian and Asselian stages from that original stage.

During last decades, the situation drastically changed, although imperceptibly for most of stratigraphers. What has happened? Suggestions concerning the methods of stabilizing nomenclature and actually indicating a return to the former nomenclature practice turned out to be inappropriate for many stratigraphers who, instead of using sound arguments, try to frighten me by saying that geological mapping will suffer disorder and collapse (Zhamoida, 1997). It is quite evident that the proposed reform of typification (introduction of hierarchy into the typification procedure) cannot itself affect the legends to geological maps. Geological mapping will undoubtedly suffer disorder and collapse, but only if we preserve the existing system of typification and allow the substantial aspect of stratigraphic scales to evolve. In this case, every substantial revision of the scale should be accompanied by revision of indices in geological maps. What, if not this, explains the existing tendency for standardization (conservation) of stratigraphic scales?

The exorbitant striving of stratigraphers to the complete fixation of the intension results in the intensional "commandments" of stratigraphic codes cannot satisfy everybody. This, in turn, results in debates, when every participant argues for the necessity to legislate his own viewpoint . Debates themselves are useful, harmful is the "fossilization" of ideas, the absolute strictness of phrasing (coding) that is peculiar of the modern official paradigm, and the incomprehension of the differences between nomenclature (technical) and terminological (substantial) aspects. For the former, the strict stability (jurisdiction) is useful and necessary, whereas for the latter, this requirement is inadmissible. Let us remember, for instance, the term "species," which is understood in different way even by biologists of the same generation, although its meaning can be understood from the context of the particular work.

Even if stratigraphers have forgotten the situation with taxonomic names prior to the 1960s, most of them know that the understanding of every fossil taxon changed significantly and repeatedly. For instance, the present meaning of the well-known genus *Spirifer* J. Sowerby, 1818, is very narrow, whereas its initial meaning corresponded to hundreds (!) of genera. Nonetheless, when reading any old article, we understand from the context what species is meant in the particular case. On the contrary, the leapfrog of new names authorized by the existing typification actually adds an element of disorder. As early as in 1989, S.V. Meyen noted that stratigraphers are literally terrorized by the appearance of new names of the Permian stages.

It is important to understand why the essentially different jurisdiction developed in biology and stratigraphy during the 1960s. I believe that there were several reasons responsible for the different fate of typification and classification in these two sciences. Most important among them are the following: the actuality of nomenclature problems in biology is much higher because of a greater number of taxa; the typological style of thinking still prevails in geology, and, as a result, an image of the stratigraphic unit is associated with the peculiar section (stratotype) in the stratigrapher's mind; practical striving for maximum legibility and stability of boundaries permanently inclined stratigraphers not only toward conventionalism and agreement on their fixation (which is quite understandable), but also toward strict fixation (by means of jurisdiction) of intensional notions and terminology that is unique and can be termed as totalitarianism in science (Lazarev, 1997). All these reasons were interrelated and favored the successful introduction of the chronostratigraphic paradigm suggesting the pragmatic approach aimed at quick and ultimate solution of problems related to the creation of stratigraphic scales.

PRAGMATIC APPROACH TO THE CREATION OF STRATIGRAPHIC SCALES: TRIUMPH OR DEADLOCK?

1954 presumably dates the appearance of the chronostratigraphic concept, which determined a new logic of the general stratigraphic scale. At that time, H. Hedberg published his report read at the session of the International Geological Congress in Alger, 1952 (Hedberg, 1954). The history of the new concept's birth was described by Hancock (1977) in a dramatic though humorous manner. Hancock considers this concept as the only significant event in stratigraphy of the 20th century. The idea was in the air, and W. Archell, a compatriot of Hancock, was also close to its "discovery," as can be judged from the essay of the latter. It should be noted that the Russian stratigrapher Stepanov (1954) concurrently introduced the term "chronostratigraphic unit" using the same meaning as Hedberg (1954), who wrote about new time-stratigraphic (chronostratigraphic) units. What idea was in the air in the 1950s and materialized almost simultaneously in the works of three stratigraphers? An idea that was grounded sound reasoning and implied that, in addition to the time calibrated by stratigraphic subdivisions known at that period, there is another time independent, of either lithological, biostratigraphic, or any other units and their successions; it is like "pure" time, within which rocks were formed. Prior to this moment, time has been considered as recorded not only in the rock successions, but also in the so-called geochronologic units, which, however, corresponded to particular stratigraphic units discernible on the basis of their specific features. Geochronologic units of that time reflected only the time aspect of stratigraphic units, and both could be considered as identical in their time ranges. In other words, all prior to the moment depended on certain subjective hypotheses (everything was based on selection of features and their interchangeability), and after 1954 time appeared to be independent of any stratigraphic features. It was a turn from subjectivity and ambiguity peculiar of our science to objectivity and

definitiveness peculiar of technological processes, i.e., the breakthrough to a new practice significantly alleviating the creating of scales, particularly of the general scale. Unfortunately, this "breakthrough" was neutral to correlation problems and even complicated them after elaboration of the scale corresponding to the measuring rule.

Chronostratigraphic units (contrary to geochronologic ones) represent rocks themselves, i.e., the rock interval limited by isochronous (ideal, imaginary) surfaces as shown in the well-known drawing by Hedberg (1965). The logic seemingly demands that no particular stratigraphic units (with real boundaries) can be considered as representing the chronostratigraphic unit beyond its stratotype or, at least, the stratotype region. However, stratigraphers do not wish to follow their own logic. Moreover, they consider it admissible that the stratotype of a stratigraphic unit is located in one region, whereas the stratotype of its boundary can be distinguished in another. It turns out that the real boundaries within the stratotype of the stratigraphic unit can never be isochronous with those in the boundary stratotype, if the latter are established in another region (in fact, no correlation methods can, by definition, base the complete isochronism).

The chronostratigraphic concept turned out to be attractive not only because of its imaginary objectivity and promise of quick and ultimate results for creation of the general scale. There was another temptation. Hedberg himself imparted the uniting role to chronostratigraphy, considering the latter as a universal key to interpretation of data obtained in all other fields of stratigraphy. As a result, many stratigraphers often do not notice the main methodological peculiarity of the new concept related to the time notion. Chronostratigraphy for them is something connected with interregional correlation or with correlation of heterofacial deposits within the same region. Moreover, chronostratigraphy for some of them became conceptually almost synonymous to the concept of ecostratigraphy or causal stratigraphy introduced by V.A. Krasilov and C.V. Meyen. For instance, this meaning is peculiar of the "chronostratigraphic subdivision" described in interesting articles by some stratigraphers (Wagner, 1993; Wagner and Winkler Prins, 1995).

The fundamental feature of the "universal key" to interpretation of different data under the causal approach is that the time itself is not a common key to different geological-historical events but, on the contrary, the common event uniting various historical-geological data into a common system (eco-, geosystem) which can represent a common (universal) key for the establishment and correlation of a time succession. Moreover, the stratigraphic unit represents our synthesized (logical) construct, i.e., biochronotype (Ruzhentsev, 1977) or, in a broader sense, stratoarchetype (Lazarev, 1997) rather than an interval between two fixed points. Correspondingly, the stratotype also cannot serve as a standard of stratigraphic unit; it represents only an approach to the name application mechanism, and the reference sections execute only an illustrative function (Lazarev, 1997). In addition, any particular section indicated in publication is an analogue of the original in biology and an analogue of the experiment in exact sciences: the case in point is the possibility of verification and falsification in the meaning of K. Popper.

The logic of appearance and development of the chronostratigraphic paradigm was as follows: (1) primarily, it was assumed that homotaxis used for correlation of different regions does not correspond to the ideal isochronism; in addition, there was a terminological discord among stratigraphers from different countries: for instance, lithostratigraphic (rock-stratigraphic) terms were often used to designate the rock ages; (2) Hedberg (1954) and Stepanov (1954) proposed the concept of chronostratigraphic subdivision as representing a totality of rocks between isochronous levels, which corresponds to a certain time interval, detectable in all the regions, where corresponding deposits exist; (3) isochronous levels bounding the chronostratigraphic one must be precisely fixed in order to be objective; (4) the situation, when not only the interval itself but also its boundaries with adjacent stratigraphic units are distinct and completely represented in the same area and, moreover, in the same section, is hardly probable; (5) accordingly, in addition to stratotypes of stratigraphic units, we must seek for other sections, even in other regions, where the boundary between adjacent stratigraphic units is detectable in the continuous succession and simultaneously distinct, even though these regions are remote; (6) inasmuch as stratigraphers did not discriminate between the nomenclature type and the intension standard both united in the stratotype concept the next question is logical: why do we need the stratorype of stratigraphic unit? It seems quite sufficient to have only the boundary stratotypes, and, thus, only one stratotype is necessary for every stratigraphic unit, namely, the stratotype of its lower boundary; (7) the intension of the stratigraphic unit itself became unimportant and was substituted by the notion of the interval between two "gold spikes," the excuse for which was the instability of the conceptual base of stratigraphic units with time; it would be simpler (more pragmatic) to reject this unstable base and consider the chrostratigraphic unit as an interval between two events; (8) the intension was yet retained as important for boundaries, and, in fact, it is desirable to select such a point in the section, which bears the best correlative potential; (9) in the last case, however, the intension was sacrificed, to a certain extent, to the gods of conventionalism and agreement because, as it is stated in the English code, none of the rock sequences, which are taken as representing some interval of the past time, can be considered as more convincing than any other rock sequence selected for the same purpose (Zhamoida *et al.*, 1969, p. 94), and this justifies

the current practice of selecting the points (boundary stratotypes) as close as possible to the previously established boundaries (hierarchy and periodization of boundaries and stratigraphic units turned out to be to be insignificant for the scale); (10) the fixed boundary stratotype can keep its position unchanged forever, unless a hiatus is proved to be in the selected point. In fact, the chronostratigraphic way of the scale development is separated from the development of the intension, and this results in the complete "fossilization" of the scale itself.

It is evident that, following Hancock (1977), many stratigraphers consider the appearance of the chronostratigraphic paradigm as an event of paramount importance and the only significant event in the stratigraphic theory of the 20th century. I agree with this assessment, but consider it to be negative rather than positive. As for positive events in stratigraphy of the 20th century, they remain less noticeable behind the prevailing paradigm. The most important ideas of the 20th century are connected with the realization of the systemic nature of stratigraphy as in works by V.A. Krasilov, I.V. Krut', and C.V. Meyen. In these works, it is clearly shown that stratigraphy should be conceptually harmonic, that all stratigraphic units are of a systemic (ecosystemic, geosystemic) nature (this means the theoretical equalizing of all the priorities of stratigraphic features), and that the time in stratigraphy is materialized and fixed in space by historical–geological processes, being unable to exist in itself. There is no room for chronostratigraphy: everything connected with the systemic approach and causality is changeable, subjective, and hypothetical, whereas chronostratigraphy claims objectivity and absolute stability, i.e., it restrains the intension.

CONCLUSION

Not long ago, prior to the 1960s, i.e., prior to the appearance of the first edition of the International Code of Zoological Nomenclature (ICZN), the issue "is there a need for the holotype?," which seems absurd for recent times, was widely discussed by biologists and paleontologists. The point is that notions "typical" and "type" were not yet discriminated by biological and stratigraphic communities at that time. Among other thins, important in the first edition of the ICZN was the clearly formulated idea on considering the nomenclature type as representing the basis of jurisdiction. Since that time, the type as a substantial notion and logical construct (an image of the stratigraphic unit) ceased to be related to nomenclature problems. This happened in biology, but, unfortunately, not in stratigraphy. Stratigraphers continue to confuse notions with the common root "type," but related to different spheres (semantic and nomenclature): "archetype–type," "typical–type"; correspondingly, other notions, such as "priority of the content" and "priority of the name," also remain indivisible.

Until recently, stratigraphers tried to make the type execute two functions at once: that of nomenclature and that of intension. This could not last forever. Chronostratigraphy solves this "schizophrenic" problem by rejecting stratigraphic classification and, as a consequence, from typification. This concept justifies the system of standards, which became a cage for the intension of stratigraphic units. Rather than guard the name, the stratotype of the stratigraphic unit turned into the procurator sentencing momentary meaningfulness to lifelong imprisonment. Moreover, after this function, the stratotype itself logically became unnecessary.

In this situation, the idea of elaboration of the International Code of the Stratigraphic Nomenclature failed and degenerated into publication of two editions of the international manual dedicated to substantial problems of stratigraphy, which is very useful for understanding of the essence of chronostratigraphy, but absolutely useless for the problem of selecting the competitive names of stratigraphic units. Only the reform in typification may provide a real basis for the creation of true jurisdiction, i.e., the real basis for the nomenclature codes, the international one included.

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Reviewers Yu.B. Gladenkov and A.I. Zhamoida

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Reviewer's remark. I think that the suggestions by S.S. Lazarev to change the procedure of typification in stratigraphy and make it similar to that in biology (to separate the name of the stratigraphic unit from the content, to consider stratigraphic units of lower rank as representing the stratotype of higher-rank subdivisions, and to divide the "stratotype" into archetype, diagnosis, and nomenclature type) will never be accepted by stratigraphers. In my opinion, this is good because such proposals can only lead to confusion, which, in turn, will result, first of all, in the paralysis of geological mapping. However, the case in point includes not only this potential consequence. Species is considered as a type of the genus because, among other things, it exists in the same time interval as the latter. However, a single bed (or member) cannot represent the type of the formation, because it corresponds, first of all, to the stratigraphic range (the time interval of the accumulation), which is quite different from that of the formation, not to mention the different contents of the whole formation and its single bed. Hence, nothing can be done in geology by ignoring the intensional aspect.

It is impossible to agree with another of Lazarev's statements that the general scale is accidental in its content and hierarchy. Surprisingly, the contingency appeared to be consistent on the planetary (!) scale. The standard is not a conditional category as stressed in the Stratigraphic Code of Russia, and boundary is characterized not only by diagnosis as follows from its definition in Russian codes, which was adopted in the second edition of the International Stratigraphic Code. The reference section is indeed original, but Lazarev did not catch its meaning. It is necessary, first of all, for the stability of stratigraphic units and elaboration of more or less reliable legends to geological maps.

The reasoning about totalitarianism is just a tribute to fashion. The uniform understanding of objects and events is one of the conditions of the fruitfulness of scientific investigations carried out by many researchers. This is an elementary matter. Codes do not "press" the solution of the intensional problems, but summarize the experience accumulated by the geological community at a given moment. Everybody can think in his own way.

I was surprised, most of all, by Lazarev's statement that the "hierarchy of stratigraphic units and position of boundaries in the general scale should change," though "spasmodically," after the disappearance of the old conceptions. This statement is accompanied by the authors' remark that the "the maturity of new conceptions is quite a subjective matter." However, life advises us to change something only in those cases when there is no possibility to leave it aside. This is true for stratigraphy as well.

A.I. Zhamoida

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Sequence Stratigraphy Workshop: Concepts and Application to Exploration and Production of Oil and Gas

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In 1996–1997, the American Association of Petroleum Geologists (AAPG) and All-Russia Research Institute of Geological Prospecting (ARIGP) organized a series of *Sequence Stratigraphy Workshops: Concepts and Application to Exploration and Production of Oil and Gas*. The workshops for specialists from CIS countries have been sponsored by the *ARCO Exploration and Production Technology* and headed by Dr. H.W. Posamentier (USA), the Foreign Member of Russian Academy of Natural Sciences, who is a leading specialist in this quickly progressing branch of stratigraphy.

The third workshop of the series was held at the State Scientific Center "*VNIIGeoinformsistemy*" in Moscow, September 14–17, 1997. Previous workshops of 1996 have been organized in Yuzhno-Sakhalinsk, Okha, and Alma-Ata.

From 50 to 70 specialists participated daily in the last workshop. They represented various industrial and research geophysical institution of Russia.

The workshop program included instructive lecturing and practical exercises, and each participant had an opportunity to ask questions and to discuss any particular subject. The participants received three methodical guidebooks: one explaining all graphical materials used by Dr. Posamentier in his lectures, and two others representing his unpublished monograph and incorporating a series of papers devoted to various aspects of sequence stratigraphy.

Dr. Posamentier is a disciple of Prof. P.R. Vail, who introduced the concept of sequence stratigraphy as a new scientific and practical branch of geology in the middle 1980s (Vail, 1987). The sequence stratigraphy analyses records of sea-level oscillations (figure) in order to divide the sedimentary cover into individual sedimentary sequences. In the opinion of Dr. Posamentier, two approaches to use the sequence stratigraphy are under development now. One, which may be termed "academical," is an objective of Prof. Vail and his colleagues. Their goal is to create a global stratigraphic scale based on eustatic sea-level fluctuations. Another "lithological" approach is adopted in professional practice, widely used by many companies in exploration of oil and gas reserves. It is based on the same concepts and has been the main subject discussed at the workshop.

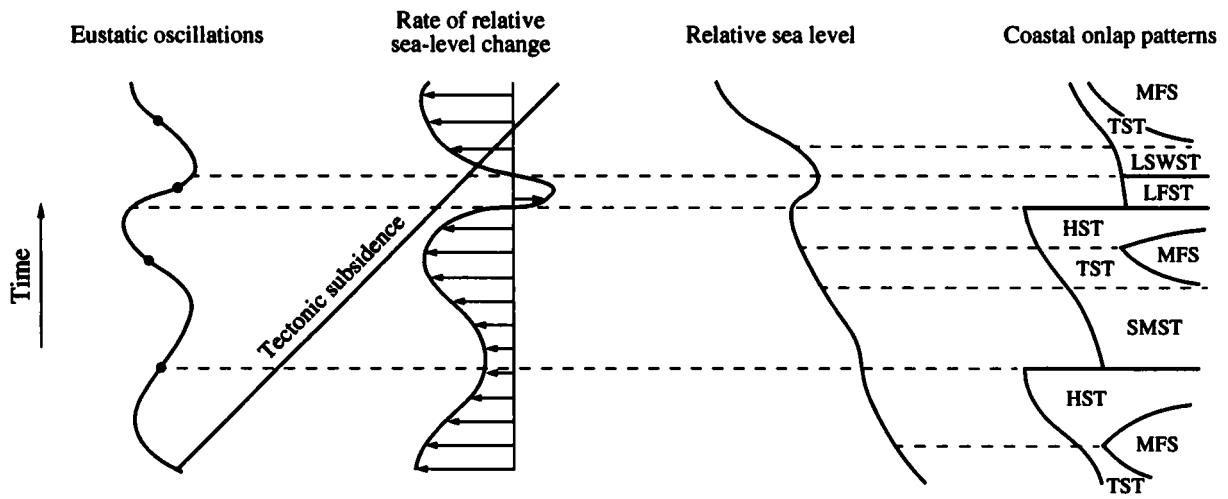
The sequence stratigraphy considers a sedimentary sequence as a stratigraphic body bounded by discordant or concordant surfaces originated in response to sea-level oscillations. A given sequence can be divided into separate units referred to as lowstand, transgressive, and highstand system tracts on the basis of data of seismic sounding and geophysical well logging (GWL) and also on the results of investigation of natural rock outcrops. The possibilities of the approach have been demonstrated using for examples some sequences depicted with the help of seismic sounding and GWL in the Gulf of Mexico and in sedimentary basins of Canada, Alaska, and China.

During the practical exercises, materials presented by the lecturer were discussed together with geophysical data on sedimentary basins studied in Russia, mostly in the West Siberian plate.

The analyzed materials exemplify sequences of system tracts variable in their time ranges and corresponding to different subsidence rates. For instance, an increase in the subsidence rate results in longer time intervals of the transgressive and highstand systems tracts, and also decreases periods of lowstand systems tracts. As each systems tract is restricted by conformity and unconformity surfaces, they open a possibility to elaborate the chronostratigraphic scheme.

The standard logging data (resistivity, self-potential, gamma-ray) also reveal different system tracts elucidating the thickness and types of sediments, their properties, and the particular significance in the sequence. The analyzed examples of sequences of regradation, aggradation, and progradation types allowed the participants of the workshop to detect system tracts of various kind and to correlate sequence of several boreholes. In order to detect the system tract boundaries more confidently, Dr. Posamentier recommended the use of the core lithology and paleontological data in addition to results of seismic record interpretation.

The main objective of the "lithological" approach in sequence stratigraphy is to predict the localization of hydrocarbon reservoirs and nonanticlinal traps for oil in a given sedimentary cover. Form this viewpoint, it is



Sea-level oscillations as a function of eustatic and tectonic subsidence. Systems tract sequences: (SMST) shelf-margin systems tract; (LFST) lowstand forced-regression systems tract; (LSWST) lowstand seaward systems tract; (TST) transgressive systems tract; (HST) highstand systems tract; (MFS) maximum flooding surface.

especially important to reveal the lowstand systems tracts corresponding to principal reservoirs. At the same time, the transgressive systems tracts are favorable for occurrences of hydrocarbon traps.

At the workshop, participants dealt with data concerning mostly the past sedimentary basins lacking signs of active synsedimentary subsidence and displaying well-detectable coastal lines, shelf zones, slopes, slope breaks, and floor areas. Basins featuring an active tectonic subsidence have not been considered. Nonetheless, the discussed methods of sequence stratigraphy can be successfully used for investigating the Upper Jurassic–Lower Cretaceous deposits of the West Siberian plate and Oligocene–lower Miocene sediments (Maikop Group) of the Scythian and Turan plates, because structures of sedimentary cover in these regions frequently reveal clinoform bedding patterns.

Some methodical aspects of sequence and seismic stratigraphy have been considered as well in publications of Russian specialists. One of monographs devoted to this problem was issued recently (Shlezinger, A.E., *Regional'naya seismostratigrafiya*, Moscow: Nauchnyi Mir, 1998). The book can be purchased from this Publishing House at the address: ul. Znamenka 11/11, Moscow, 119890 Russia; tel. (095) 291-2847.

Reviewer Yu.B. Gladenkov

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